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A sedimentological study of the glacial deposits at
Mohawk Bay, near Dunnville, Ontario.

by

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of the requirements for the degree of
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Geological field 1985

Abstract

Detailed lithofacies analysis of the late Pleistocene deposits at Mohawk Bay (near Dunnville, Ontario) suggest that these sediments are of subglacial and glaciolacustrine origin. The deposits form a fan-shaped ridge with the long axis parallel to regional ice movement. The sediments prograde vertically from a basal lithofacies assemblage that records a tripartite subglacial stratigraphy into waterlain diamicts and sands, that form two distinct stratigraphic units.

The lowermost unit (unit A) was deposited initially by lodgement processes, superseded by development of a subglacial meltout sequence, followed again by lodgement. Meltout was initiated by the freezing of subglacial melt-water sediments which resulted in an increase in basal debris shear strength, effectively causing the basal ice to stagnate. The overriding ice deposited lodgement till and resulted in deformation of the stagnant debris/ice mixture.

The overlying units record deposition in the proximal

zone of a proglacial lake, in front of a calving ice wall. Unit B, exposed along the flanks of the ridge is composed of two diamict lithofacies which were deposited by cohesive debris flow and/or suspension fallout. A sand lithofacies is also present which was the result of turbulent mass flows and/or suspension fallout. Unit C, present at the centre of the ridge, was deposited as a subaqueous outwash fan. Predominantly it is composed of sandy, thin-bedded turbidites, cohesive debris flows, and channel-fill sequences. Soft-sediment deformation as the result of water escape is extensive and has caused the destruction of the majority of primary structures and internal stratigraphy. Together these two units comprise a morainal bank which marks the position of a very brief stagnation period of the retreating ice lobe.

It is hypothesized that the upper lodgement sub-facies unit A was deposited during a surge that may have existed just prior to deglaciation. Increased meltwaters would have resulted in saturation of the basal debris, lowering basal frictional component, thus allowing velocity to increase substantially. This increased

velocity would also have the effect of increasing tensile stresses on the ice. Calving rates would increase because of these stresses, resulting in a calving wall.

Stagnation of this calving ice wall in the vicinity of Mohawk Bay is responsible for deposition of the upper stratigraphic units. Orientation of the ice front would have been approximately perpendicular to the long axis of the ridge with the material comprising the morainal bank emanating from a subglacial conduit. It is suspected that sedimentation occurred very rapidly, perhaps in as little time as a few months.

The proposals of this work suggest the re-examination of late Pleistocene landforms in the Niagara Peninsula and their relationship to deglaciation.

To my family: my parents, Francis Matthew and Audrey, and my sister, Christine Elaine (McKay), without whose encouragement and support this thesis would have remained a figment of my imagination.

Acknowledgments

As is the case with most theses, a number of individuals assisted the author in the preparation of this work. First and foremost, it is a pleasure to thank Dr. John Menzies, who was principal supervisor. Throughout numerous discussions Dr. Menzies provided constructive criticism, insight and enthusiasm for the subject of glacial geology in general. Dr. Howard Williams made useful comments and suggestions during the project and it is from him that I learned the importance of detailed field work. Dr. Jaan Terasmae also provided informative comments on various aspects of the thesis.

Field work during the long, hot summer of 1983 was conducted with the assistance of my good friend and Taoist philosopher, John Balinski. Without his help, I doubt whether the field work would have been as productive or enjoyable as it was. Ian Wilson and Paul Fredrick also assisted in the field and laboratory. Peter Brown's drafting excellence was put to good use in figure 2.1.

Finally I would like to thank my fellow graduate students, the Geol. 552 scholars. These people, who are too numerous to thank individually, provided little in the way of scientific input to the thesis, but were a lot of laughs and made graduate studies at Brock a hell of a good time.

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Chapter 1

Introduction

1.1 General

Research into aspects of glacial sediments has been on-going since the mid-1800's. The type of research has been far reaching: stratigraphy and geochronology (e.g. Terasmae, 1979, 1980), extinction and vegetation history (e.g. Winn, 1978), geotechnical applications (e.g. Boulton and Paul, 1976; Eyles and Sladen, 1981), and sedimentological studies with the emphasis on facies modelling (e.g. Eyles and Miall, 1984; Powell, 1984).

Sedimentological studies have tended to be oriented towards laboratory methods: grain size analysis, heavy mineral and carbonate contents, and clast lithology and mineralogy (e.g. Gravenor, 1980). Furthermore, these parameters have been used for stratigraphic purpose in order to correlate glacial deposits from different geographic areas (Karrow, 1984).

An entirely different approach has recently been gaining favour amongst workers in glacial geology. This has been the utilization of physical sedimentological characteristics. Features like primary and secondary structures, bedding contacts, and facies associations, commonly studied in the non-glacial environment, have been examined to determine processes responsible for deposition. Also, research into modern glacial environment has revealed a wealth of material, of which analogues in ancient sediments have been identified.

By using both field and laboratory methods, processes and environments can be determined, leading to the development of suitable depositional models.

1.2

Present study

The present study involves the facies analysis of a sequence of late Pleistocene glacigenic sediments exposed along the north shore of Lake Erie, southeast

of Dunnville, Ontario, at Mohawk Bay (fig. 1.1). The deposit forms a fan-shaped ridge with the long axis trending northeast-southwest. Outcrops are located along the shoreline which is affected by present-day slumping; because of this exposures tend to be altered, sometimes drastically.

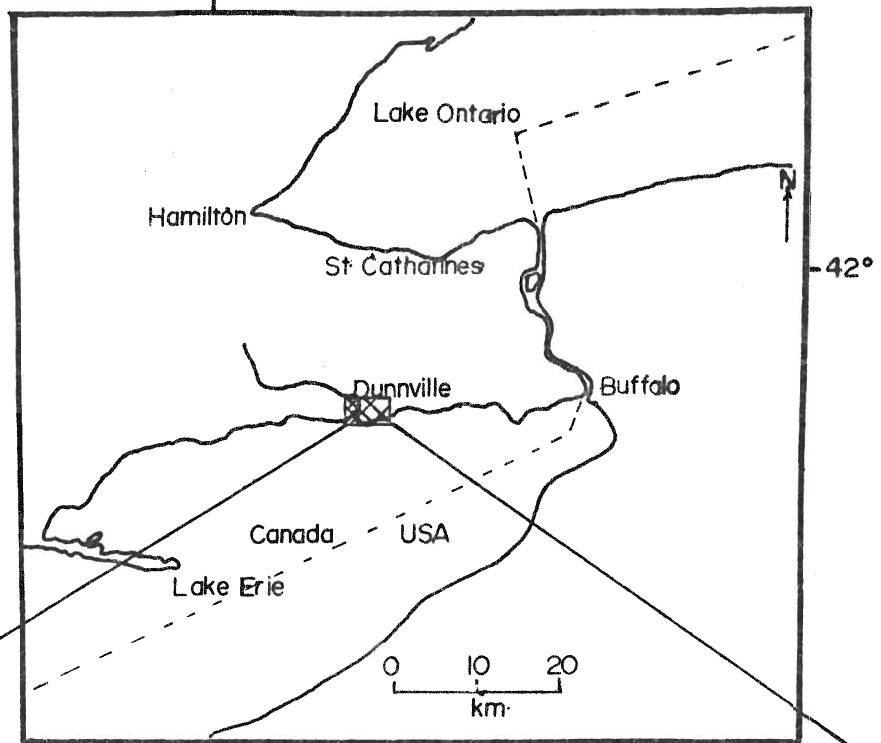
Previously, Feenstra (1974, 1982) stated that this feature was a recessional moraine that marked a stagnant ice front position during retreat of the Ontario-Erie lobe from the Niagara Peninsula. Feenstra (1974) termed the feature the Port Maitland Moraine.

Since Feenstra (1974, 1982) briefly studied this feature as part of a reconnaissance project of the Niagara Peninsula Quaternary sediments, it was thought that a detailed lithofacies analysis should be conducted to either confirm or dispute his interpretations.

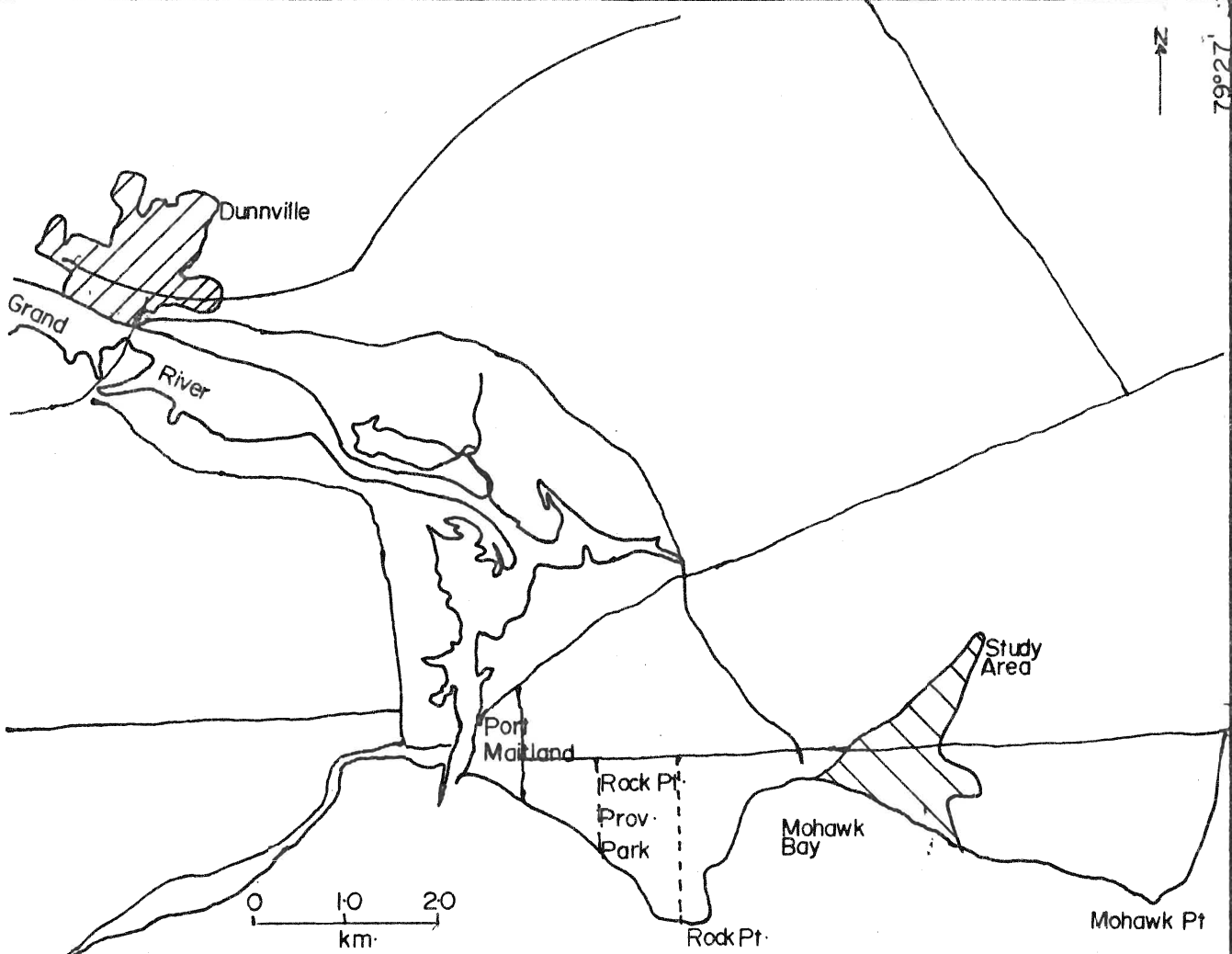
Therefore, the purpose of the thesis is four-fold:

- 1) to determine the sedimentological and lithofacies relationships of the feature;
- 2) from these relationships, ascertain the processes and environments of deposition;

Figure 1.1. Location of study.



42°57'



- 3) to formulate a depositional model;
- 4) to relate these findings to the glaciation/deglaciation history of the Niagara Peninsula.

1.3

Previous work

Work on the sediments at Mohawk Bay has been conducted in general fashion by Chapman and Putnam (1966) and St. Jacques and Rukavina (1973). More detailed work followed, by Feenstra (1974, 1982) and Fordham (1981). All of these researchers refer to the features as the Port Waitland Moraine. Both Feenstra (1974, 1982) and Fordham (1981) speculate that the sediments were formed at an ice front that was calving into a proglacial lake, but never explored the sediment-ice dynamics relationships.

As was mentioned earlier, the field of glacial and Quaternary studies is widespread. In the author's opinion, the best overview of glacial processes and

sediments is that of Sugden and John (1976) and Ashley et al., (1985), to which the reader is directed for any general information.

1.4

General geology

1.4.1

Bedrock

The Niagara Peninsula bedrock is composed of virtually flat-lying sediments of Ordovician to lower Devonian age (fig. 1.2; table 1.1; Rickard and Fisher, 1970). These rocks form two north-facing escarpments: the Niagara in the north and the Onondaga in the south. The former represents the major topographic feature in the area, the latter is almost covered by Quaternary sediments and can only be traced through borehole measurements.

Figure 1.2 Bedrock geology of the
Niagara Peninsula.

- 1) Queenston Formation
- 2) Whirlpool, Power Glen, Thorold,
Grimsby formations
- 3) Reynales, Neagha, Irondequoit
formations
- 4) Rochester and Decew formations
- 5) Lockport Formation
- 6) Salina and Bertie formations
- 7) Bois Blanc Formation
- 8) Onondaga Formation

(after Rickard and
Fisher, 1970)

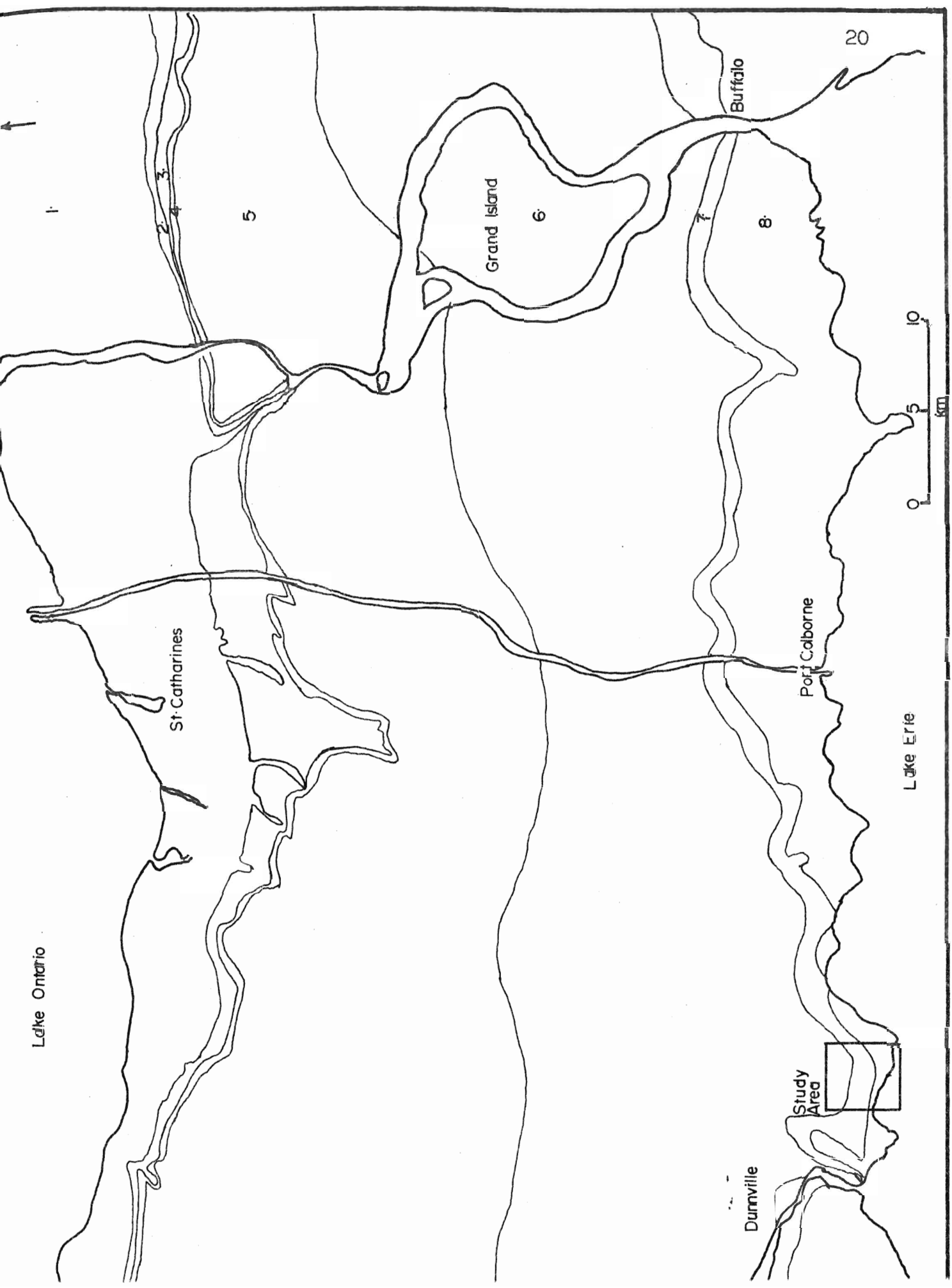


Table 1.1

Bedrock Lithologies of the Niagara Peninsula

Age	Formation	Member	Description
Lower Devonian	Onandaga	-Clarence	-massive, cherty limestone
		Edgecliff	-medium bedded, bioclastic limestone
	Bois Blanc		-medium bedded, cherty, bioclastic limestone.
	Bertie		-bituminous, argillaceous dolomite
Upper Silurian	Salina		-argillaceous dolomite, shale and evaporites
	Guelph	Eramosa	-massive dolomite
	Lockport	-Coat Island	-bituminous dolomite
		Gasport	-massive, fine grained argillaceous dolomite
	Decew		-crinoidal, dolomitic limestone
			-argillaceous to arenaceous dolomitic limestone
	Rochester		-calcareous shale to siltstone
	Irondequoit		-crinoidal limestone
	Reynales		-massive, dolomitic limestone
Lower Silurian	Neahga		-shale, minor limestone
	Thorold		-massive, quartzose sandstone
	Grimsby		-crossbedded sandstone and shale
	Power Glen		-shale, calcareous sandstone
Ordovician	Whirlpool		-crossbedded, quartzose sandstone
	Queenston		-red shale

(after Bolton, 1957; Rickard and Fisher, 1970; Tarrant and Telford, 1975).

• The field area is underlain by Onandaga, Bois Blanc, and Bertie formations, all of lower Devonian age. The Onandaga and Bois Blanc formations are cherty limestones. The Onandaga is further subdivided into the Clarence and Edgecliff members which differ by way of chert and fossil content and the amount of intercalated argillaceous material. The Bertie Formation is a bituminous dolomite with minor argillaceous, mottled and laminated dolomite interbeds (Tarrant and Telford, 1975).

1.4.2

Quaternary

Records of the glaciation in the Niagara Peninsula date to at least 23000 yrs BP (Karrow and Terasmae, 1970). The majority of the deposits in the area, however, are of late Wisconsinan age (Feenstra, 1982) and are related to the last advance/retreat of the ice through the region. The till sheet that blankets the underlying bedrock is referred to as the Walton Till (Terasmae, et al., 1972)

and was deposited about 13000 years BP (Terasmae and Matthews, 1980). The ice lobe that was responsible for this advance was the Ontario-Erie lobe that advanced into the Lake Erie basin while Lake Whittlesey was in existence in the basin (Prest, 1969; Barnett, 1979). The extent of the ice is given by the positions of the Paris and Galt moraines (Winn, 1978). This event occurred during the Port Huron Stadial (Dreimanis, 1977).

Subsequent to this advance, interstadial conditions ensued during the Two Creekian Interstade. In the Niagara Peninsula, this event is marked by the development of the Erie Basin glacial and post-glacial lake phases and the formation of Lake Iroquois in the Lake Ontario basin (Fullerton, 1980).

The deglaciation during the Two Creeks Interstadial in this region is illustrated, according to Feenstra, (1982), by the formation of several "recessional moraines". These are the Port Maitland, Crystal Beach, Fort Erie, Niagara Falls and Vinemount moraines which supposedly mark the ice front position as the Ontario-Erie lobe retreated.

Other deglaciation features associated with this retreat are the Fonthill kame/delta, the Short Hills complex, and the Lake Iroquois shoreline, all found in the northern part of the Niagara Peninsula (Feenstra, 1982; J.Z. Fraser, personal communication, 1983). These features are illustrated in figure 1.3.

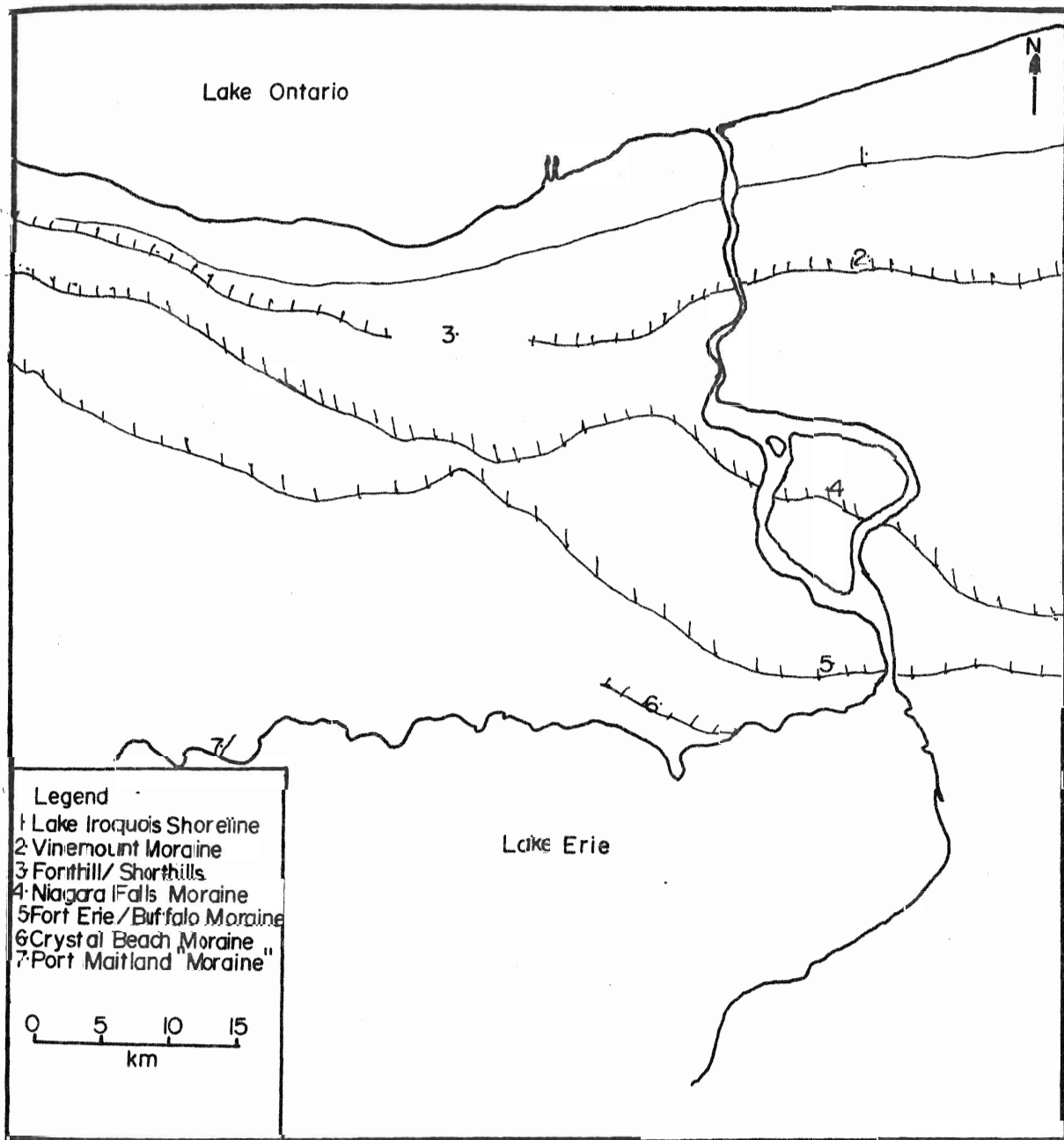


Figure 1.3. Late Pleistocene landforms of the Niagara Peninsula (after Muller, 1977).

Chapter 2

Field Studies

2.1 Introductory Remarks

In order to ascertain the depositional history of the sequence exposed at Mohawk Bay, a two-fold approach was adopted: the main thrust to be based on field observations, later to be supplemented by laboratory analysis. Field work involved the location of appropriate sections, followed by measurements and detailed descriptions of the sediments. Accessibility was a problem at times; therefore estimations of measurements were made occasionally. Descriptions entailed making note of such features as primary and secondary structures, bed contacts and bed geometry, and facies associations.

To aid in describing the sediments, the coding system of Miall (1977, 1978) and Eyles et al. (1983) was employed. Similar coding schemes have recently been applied to glacial sediments, with success, by other workers, notably Domack (1983) and McCabe et al. (1984). Both have used codes based on Miall (1977; 1978) and Eyles et al. (1983). This scheme is presented in table 2.1.

Laboratory analysis entailed grain size determinations. Wet sieving was used to separate the $< 4.0\phi$ and was followed by dry sieving at 0.5ϕ intervals.

Table 2.1

Lithofacies Code System (after Miall, 1977, 1978; Eyles et.al., 1983).

Diamict - D

Dm	:	matrix supported
Dc	:	clast supported
D - m	:	massive
D - s	:	stratified
D - g	:	graded
D---(r)	:	resedimented
D---(c)	:	current reworked
D---(s)	:	sheared

Gravel - G

Gms	:	massive, matrix supported
Gm	:	massive to crudely bedded
Gt	:	trough crossbedded
Gp	:	planar crossbedded

Sands - S

Sr	:	rippled
St	:	trough crossbedded
Sh	:	horizontal lamination
Sm	:	massive
Sg	:	graded
Sd	:	soft-sediment deformation

Fine-grained (silt/clay) - F

F1	:	laminated
Fm	:	massive
F - d	:	with dropstones

2.2 Stratigraphy

Detailed mapping of the section revealed that three major stratigraphic units (A, B and C) could be identified. Figure 2.1 illustrates the stratigraphy and details of the individual logs.

2.2.1 Unit A

Initially, this unit was mapped as two distinct lithological units. However, it was decided that in order to simplify matters, the two units could be grouped together as one kineto-stratigraphic unit, following the work of Berthelsen (1978). The idea of kineto-stratigraphy, which has been applied to multi-glaciated terrains showing extensive glaciotectionism, is that the base of a unit may be defined by the limit of penetrative deformation. It will be shown later that these two units have undergone deformation by the same processes, thus it seems appropriate that they should be grouped together.

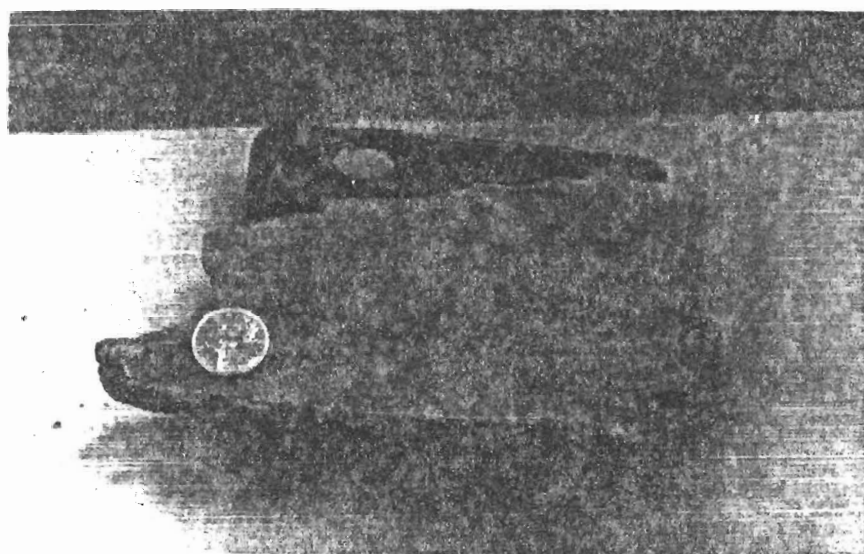
The lower lithological unit, Dmss, is exposed over a very small area (fig. 2.1). Thickness of the unit does not exceed 2.0m above beach level. From this limited exposure, it was found that Dmss is a fine grained, stratified sediment with virtually no clasts (fig 2.2). Figure 2.2 (b) shows distorted and apparently attenuated

Figure 2.2. Dmss. (A) Dmss with horizontal layering.

(B) distorted and apparently
attenuated layering

In both (A) and (B) note the lack
of clasts and the fine-grained
texture of this diamict.
2.2cm coin for scale.

A



B

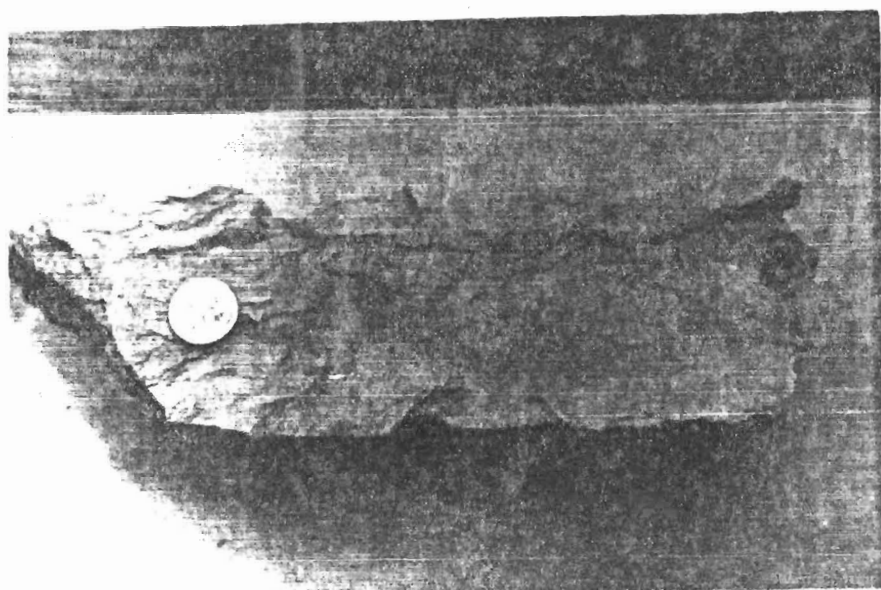
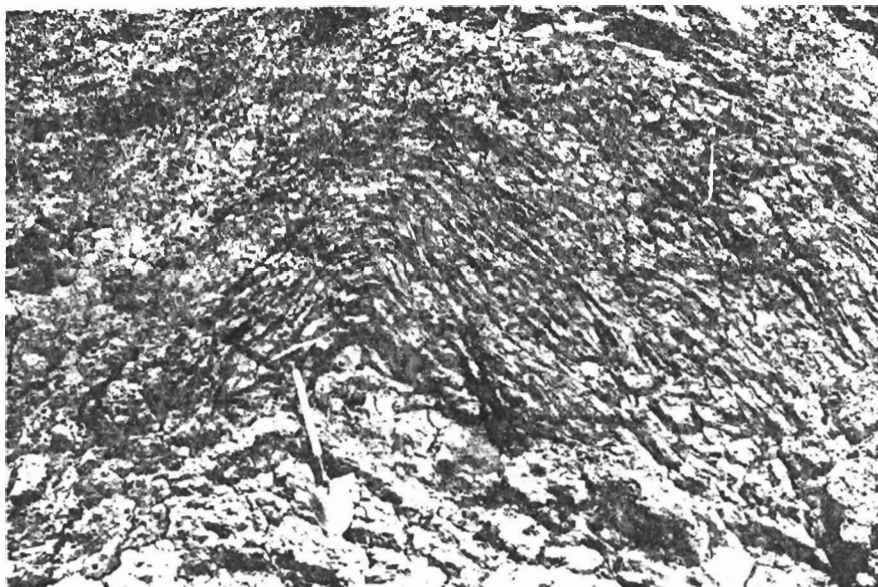


Figure 2.3. Field exposures of Dmss.

- (A) anticlinal structure within Dmss. Sediment found within this structure is the highly attenuated variety shown in figure 2.2 (B).
- (B) Discontinuity within Dmss. Close examination of the photograph reveals a subvertical layering that has been truncated by the overlying sediment.
15cm ruler for scale.

A



B

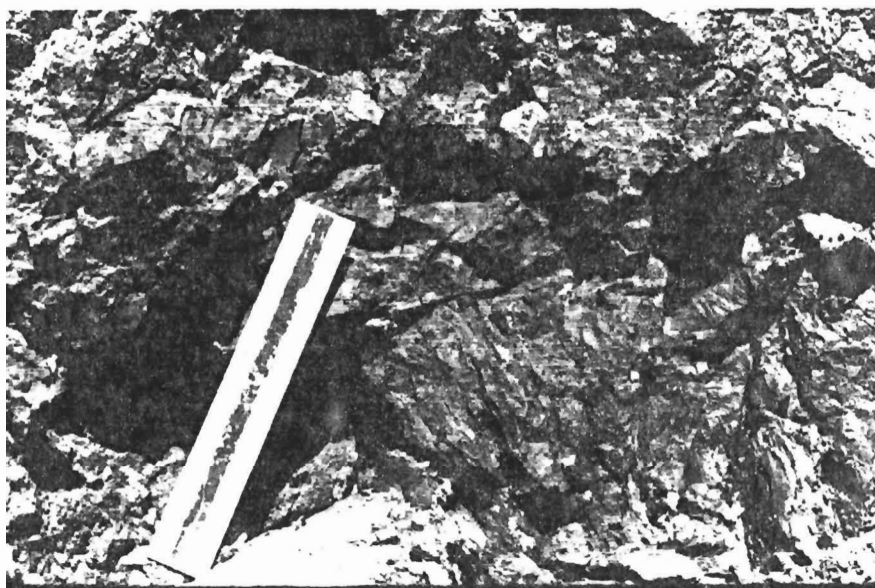
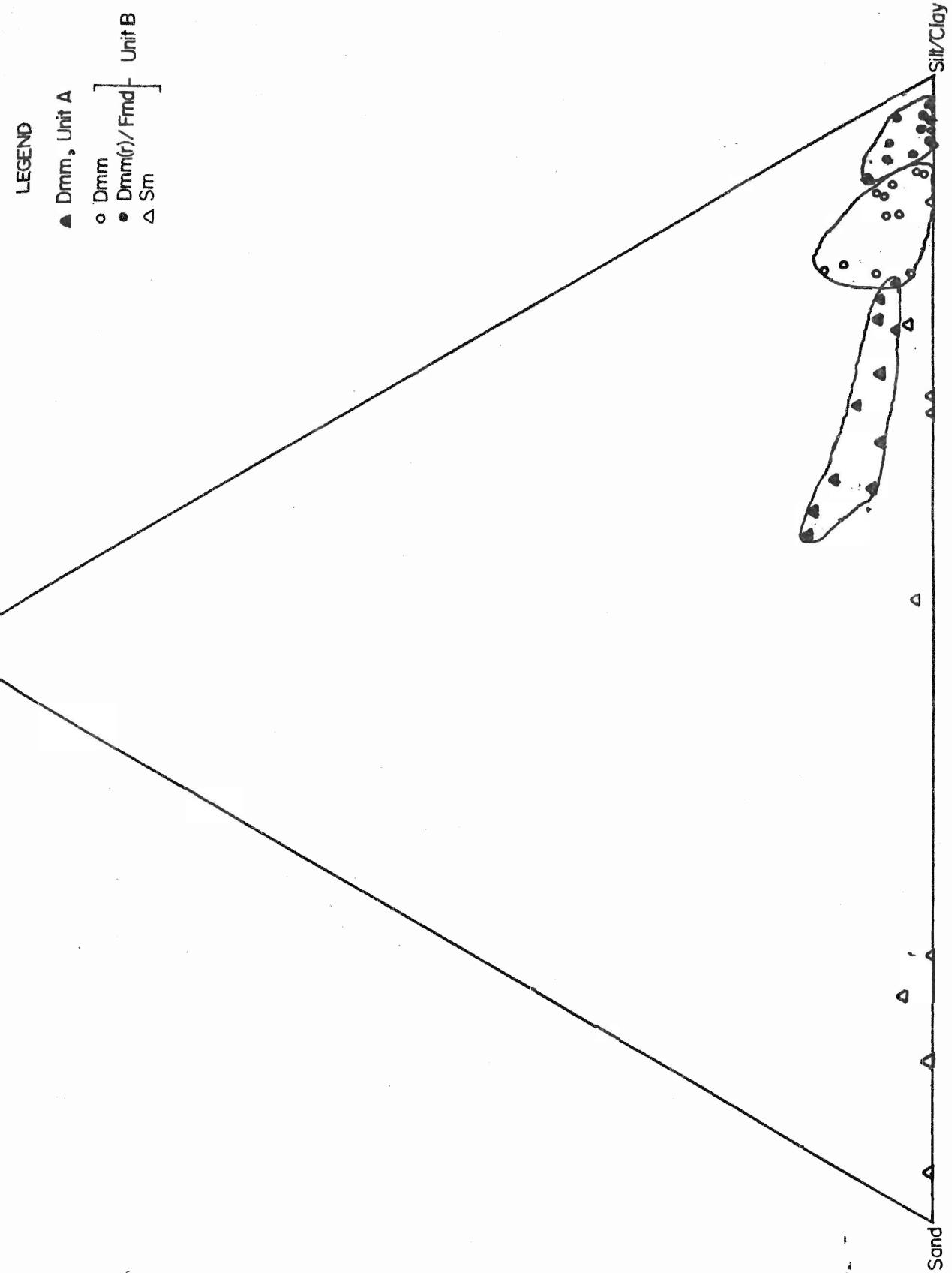
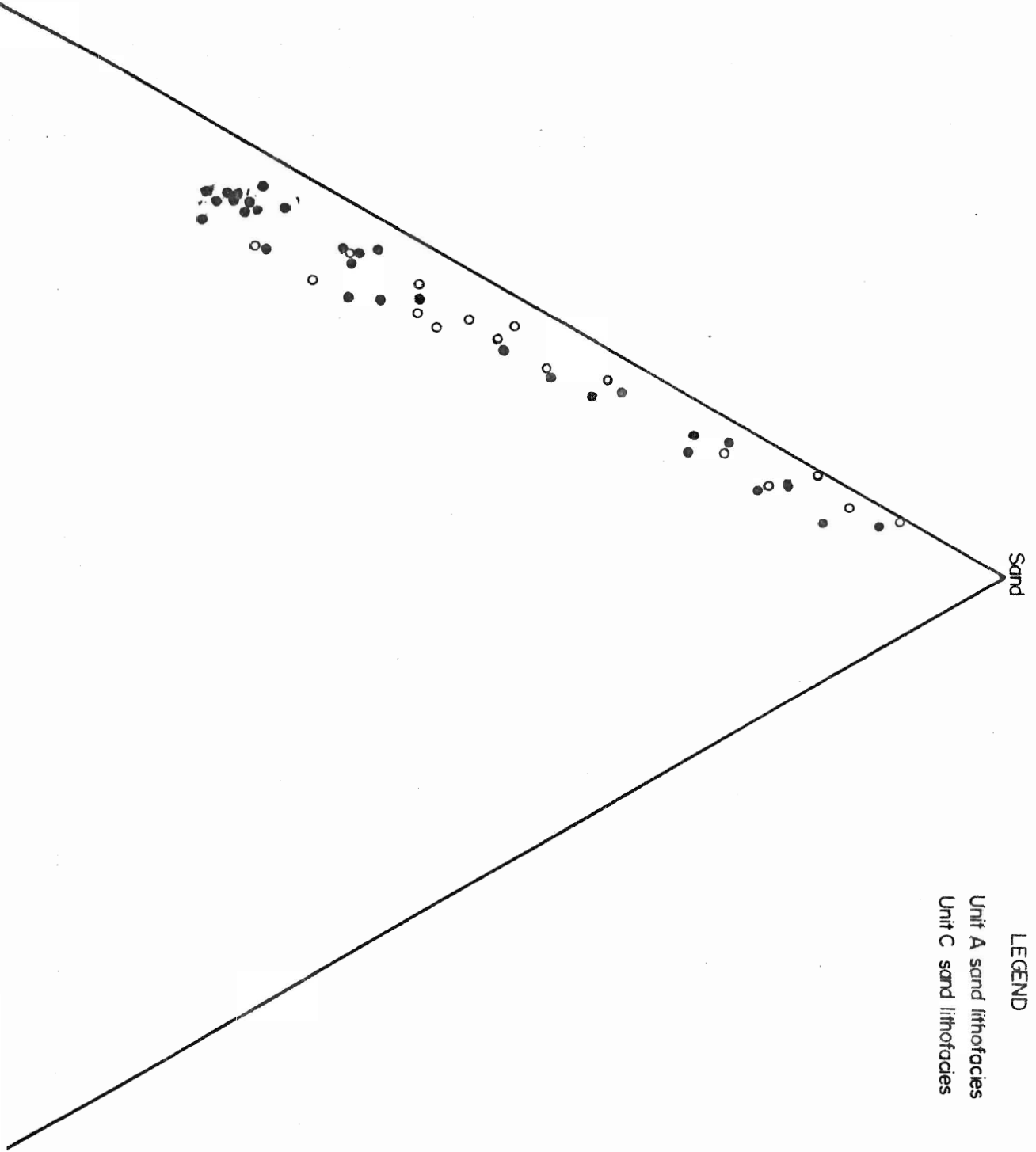


Figure 2.4. Ternary plots of grain size data.



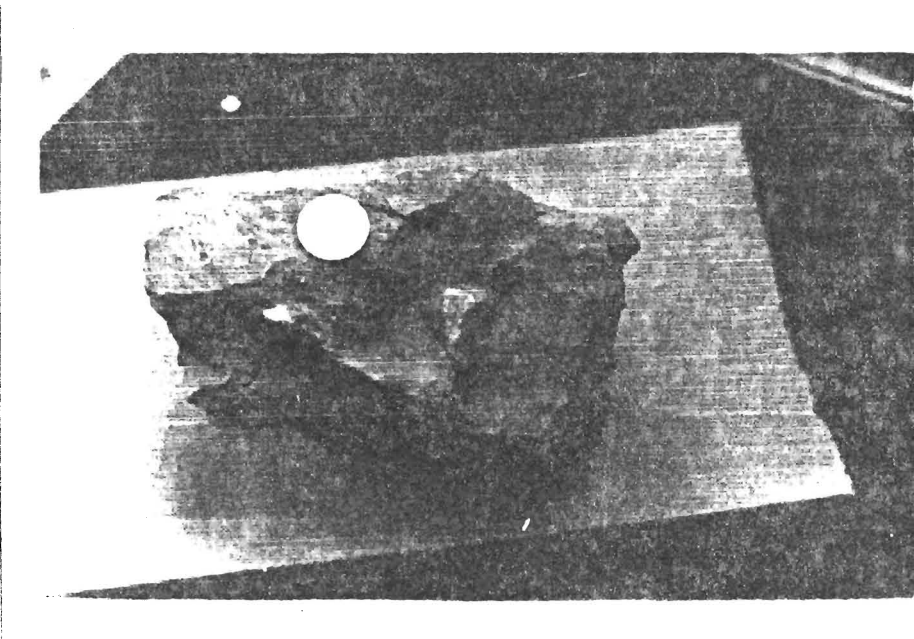


- Figure 2.5. (A) Dmm exposed at the base of the section. Note the presence of vertical and sub-horizontal fissures, and moderately stoney appearance. Increase in clast concentration at the base is due to present-day wave action. Pick for scale.
- (B) Close-up of Dmm. Matrix has been deformed around clast.

A



B



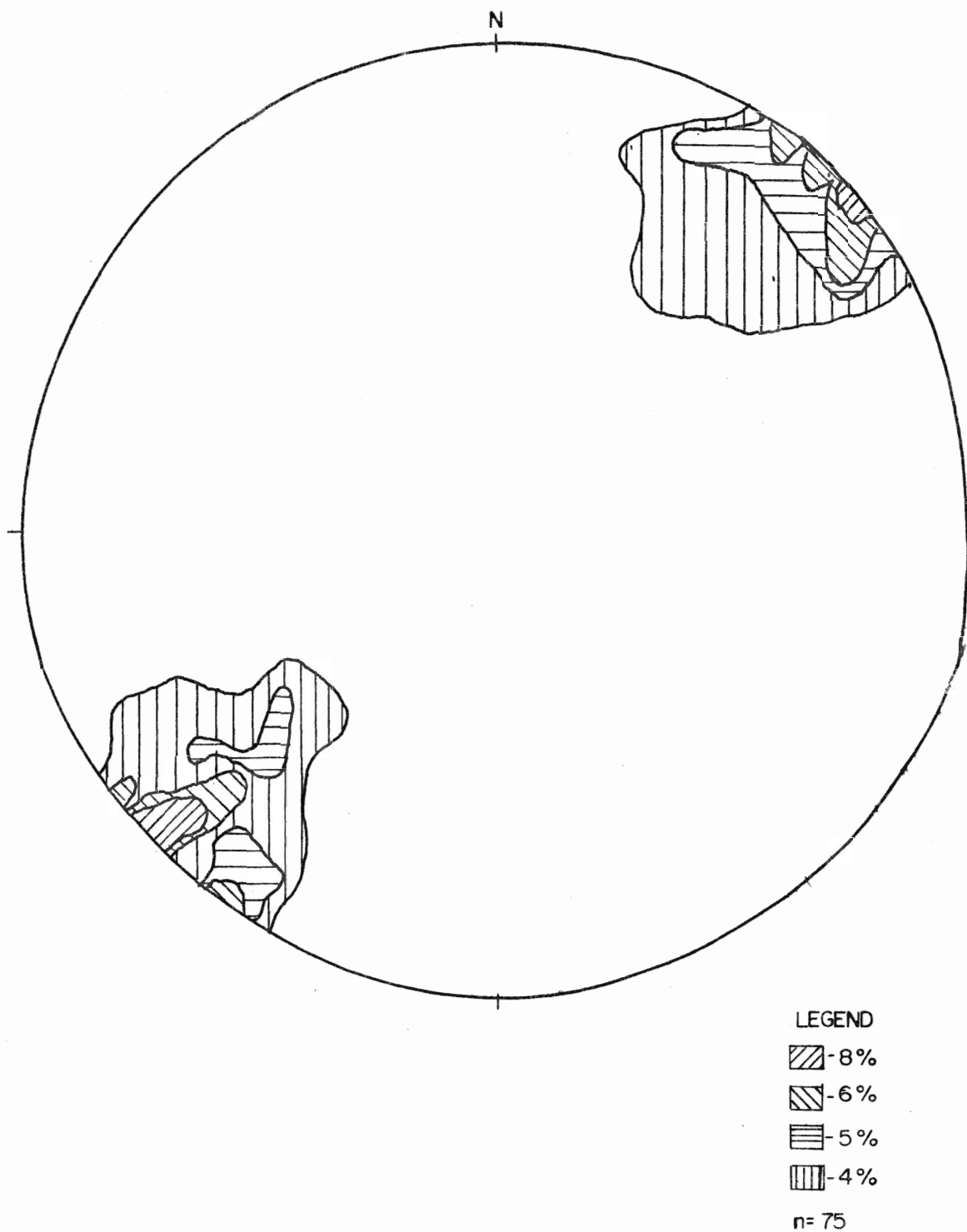


Figure 2.6. Equal-area projection of fabric

layering. On a larger scale, folding of Dmss is present (fig. 2.3 (a)). The contact with the overlying lithological unit is sharp and conformable, in this exposure, as depicted in figure 2.3 (b).

The upper lithological unit is composed of one diamict lithofacies. Within the unit are several sand lithofacies which are found as a number of discontinuous, podiform to lens-shaped inclusions (fig. 2.1). The thickness of the unit ranges from 7.0 to 14.0m, with the diamict lithofacies, Dmm, being exposed continuously throughout the entire length of the section. It is a low to moderately stoney, highly compacted diamict with a silt clay matrix (fig. 2.4, 2.5). Clasts appear to be predominantly of local origin, reflecting the Palaeozoic bedrock of the southern Niagara Peninsula, are subrounded to sub-angular and striated. Fabric measurements (fig. 2.6) indicate a predominance of a northeast - southwest trend with shallow plunges. Clasts were found to occur in clusters (cf. Boulton, 1975; Kruger, 1979): these may have formed when clasts concentrate along the stoss side of another clast.

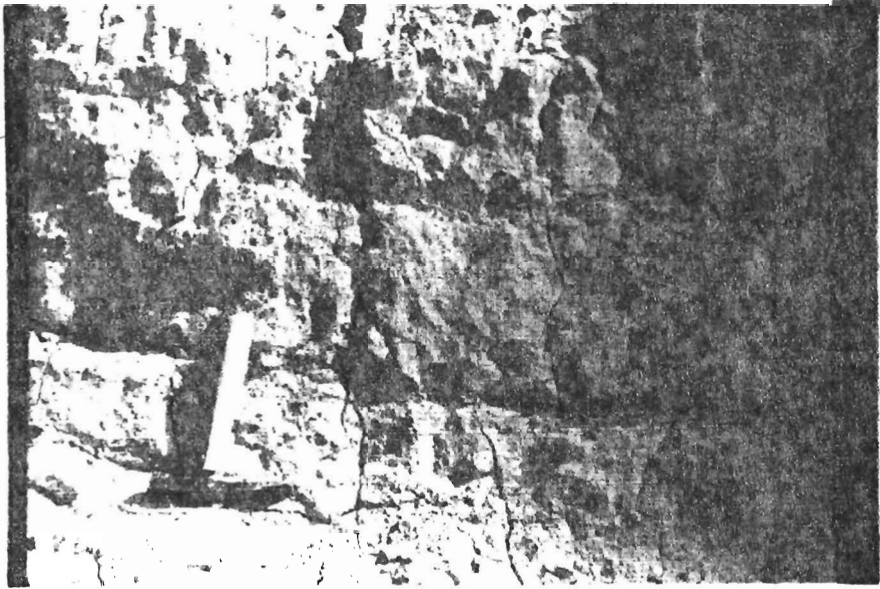
Another feature found in Dmm, are zones where disaggregation (crushing?) of the coarse fraction of the diamict has occurred (fig. 2.7). These zones are thin (<1.0cm) and may be continuous over about 5.0m. Sand lenses associated with these zones appear to have been

Figure 2.7. Shear zones within Dmm.

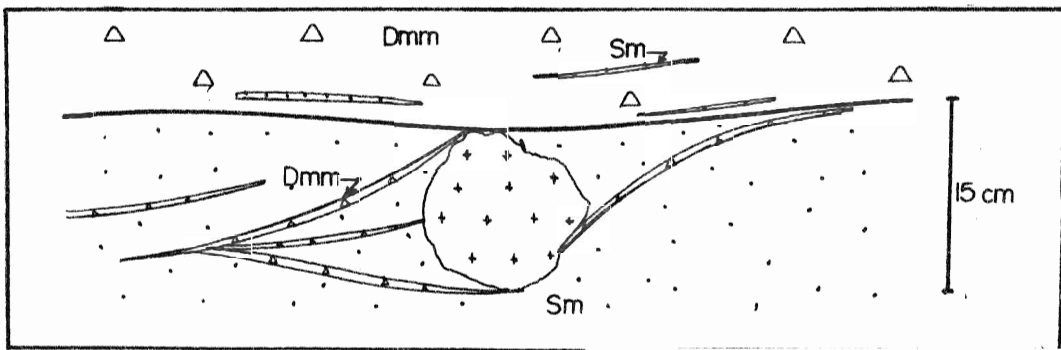
(A) Subhorizontal shears within Dmm. Close examination of these features showed disaggregation of coarser material. Hammer for scale.

(B) Field sketch of a shear zone at a Dmm/Sm contact. Note deformation of Dmm layers around clast and stringers of Sm that appear to have been attenuated. Lineation directions are oriented obliquely out of the page.

A



B



attenuated and deform around clasts (fig. 2.7 (b)). Eyles and Sladen (1981) and Rappol (1983) have described similar features in tills in northern England and West Germany, respectively, and have concluded that they are resultant of a shearing process. Also Boulton et al. (1974) have documented crushing to be operating under conditions of subglacial shear. Thus these zones within Dmm shall be referred to as shear zones.

The most striking phenomenon of this unit is the occurrence of the intra-diamict sand facies. These units are 2.0m to 19.0m across and are generally ≤ 4.5 m in thickness (fig. 2.8). Grain size analysis shows that these are primarily silty sands (fig. 2.4). As can be seen from figure 2.8, certain inclusions appear to exhibit macro-scale deformation. Those deposited in figures 2.8 (d) and (e) are apparently folded and the upper inclusion in figure 2.8 (f) appears to have undergone rotational movement. Inclusions in figure 2.8 (a) and (b) seemed to have escaped much larger scale deformation.

The amount of macro-scale deformation is supported by the fact that smaller-scale deformation increases correspondingly. The least deformed blocks in figures 2.8 (a) and (b) have well-preserved primary structures. Figure 2.9 (a) and (b) show the Sr facies of the inclusion in figure 2.8 (a). Both type A (erosional-stoss) and type B (depositional stoss) climbing ripples (Allen, 1973) are present.

Figure 2.8. Intra-diamict sand facies.

(A) and (B). Inclusions which have apparently escaped macro-scale deformation. Note the lens-to-lenticular--shaped geometry of these inclusions.

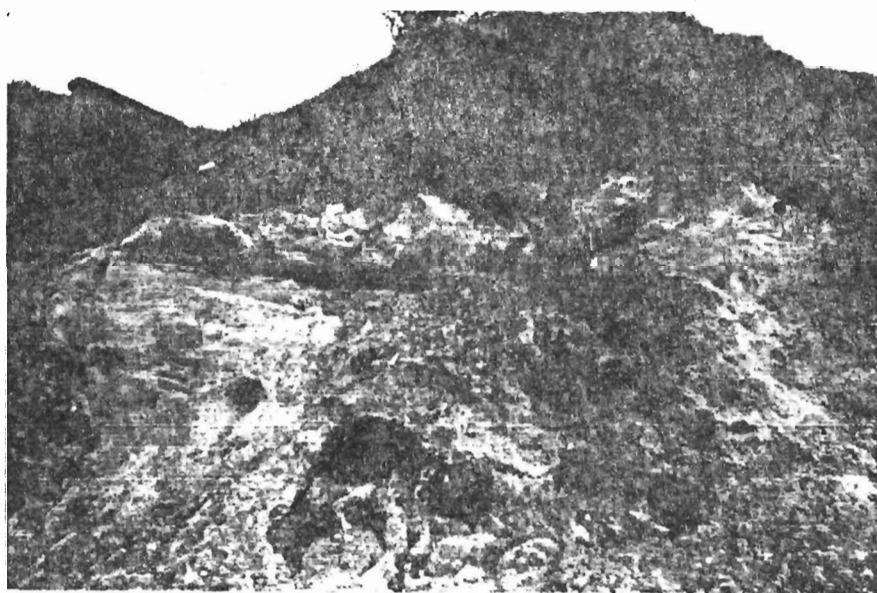
(C) to (E). Increasing macro-scale deformation.

(C) Sand block showing tilting and diamict injection.

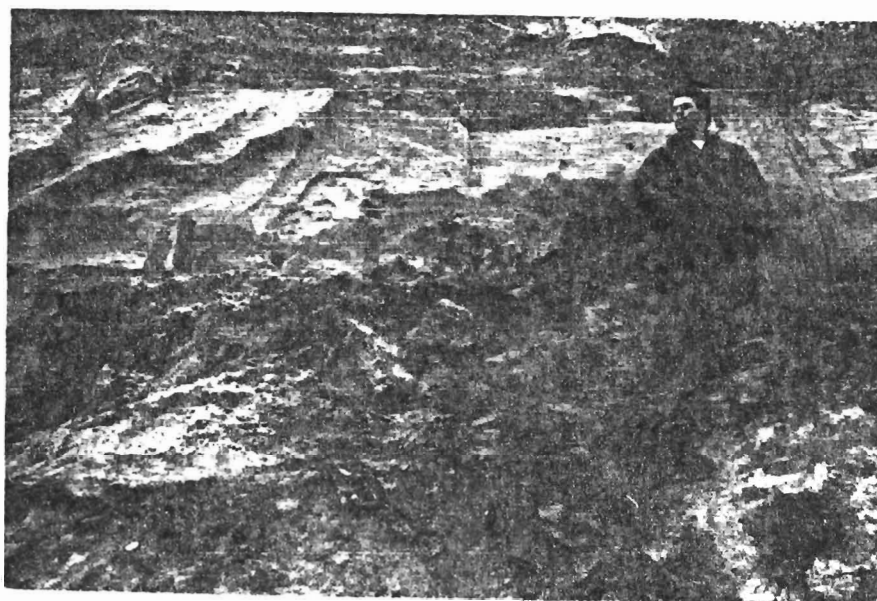
(D) and (E). Folded inclusions.

(F) Sands showing intense deformation. Upper inclusion appears to have undergone rotation and diamict injection. Close examination of this section revealed the presence of a shear zone overlying these inclusions.

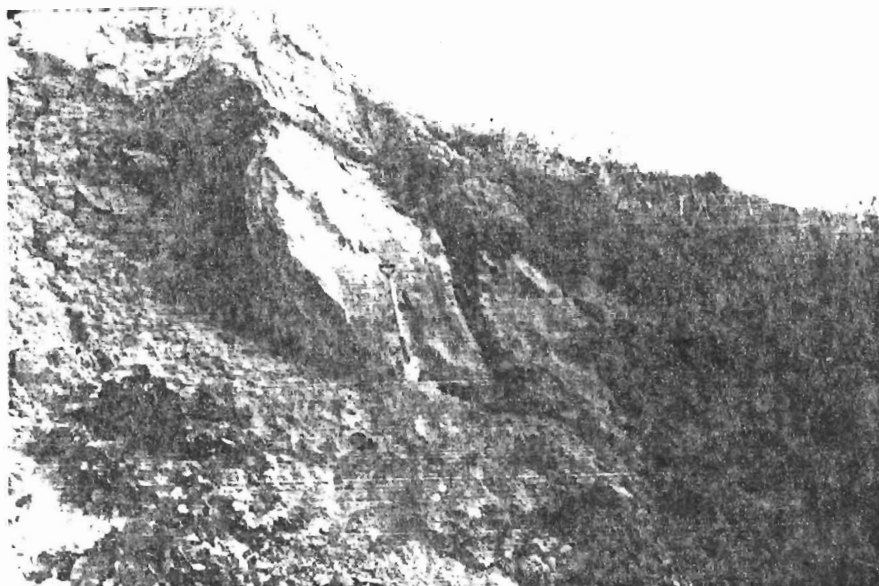
A



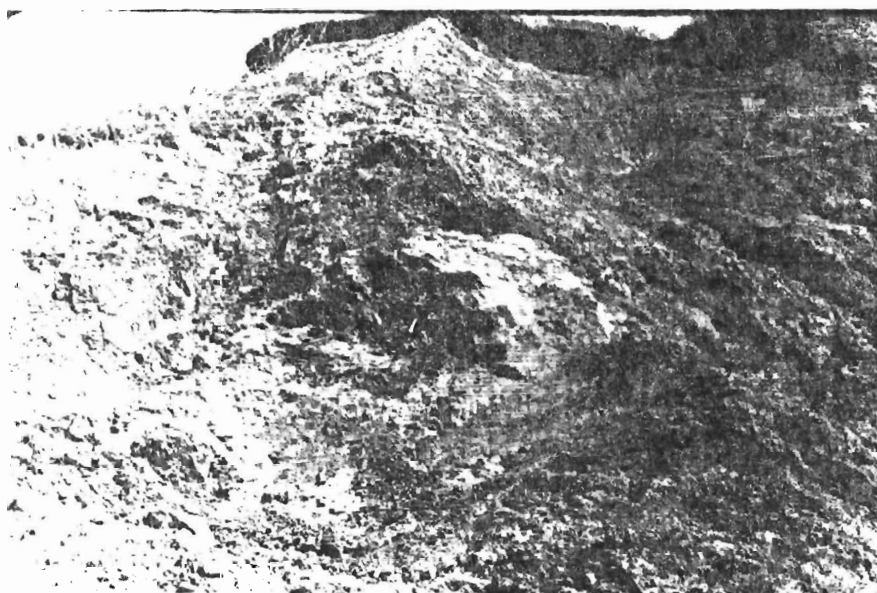
B



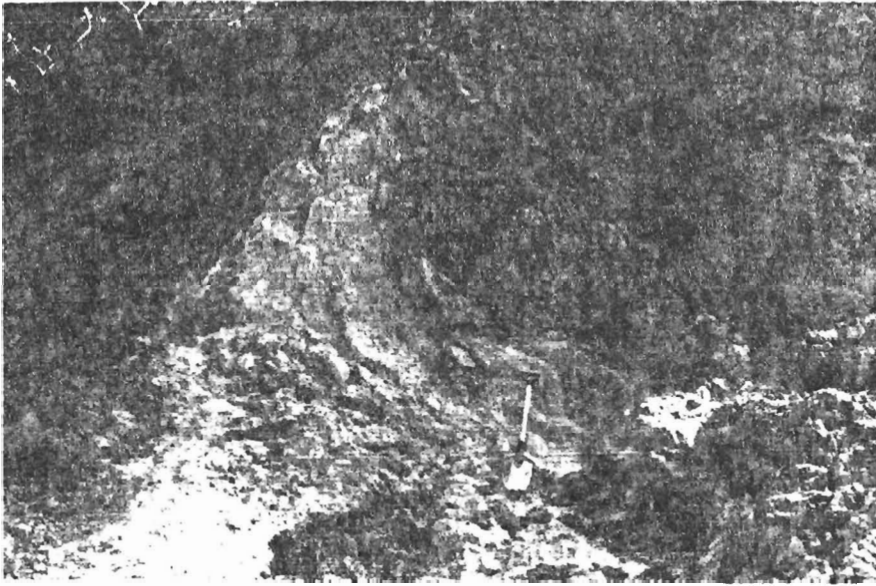
C



D



E



F

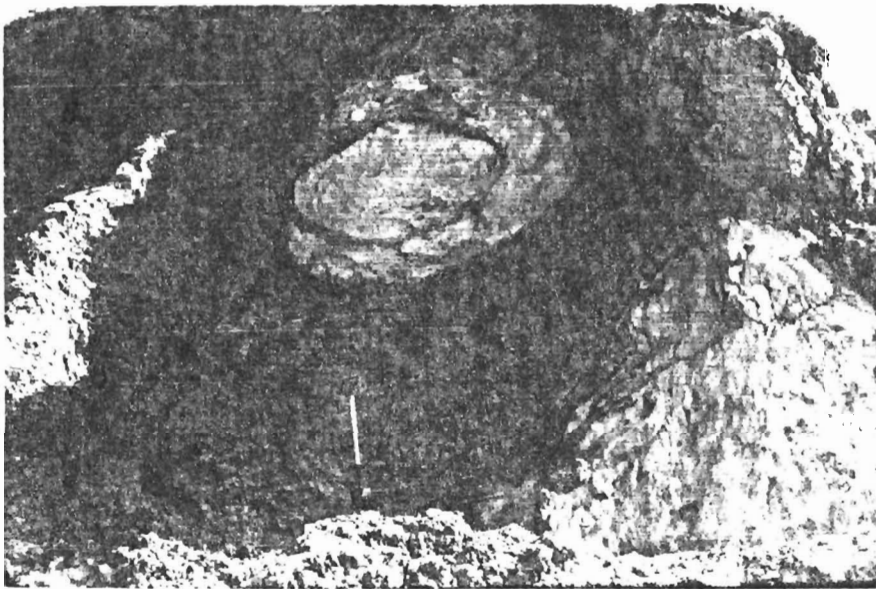


Figure 2.9. Preserved primary structures in sand inclusions. Sr lithofacies illustrated.

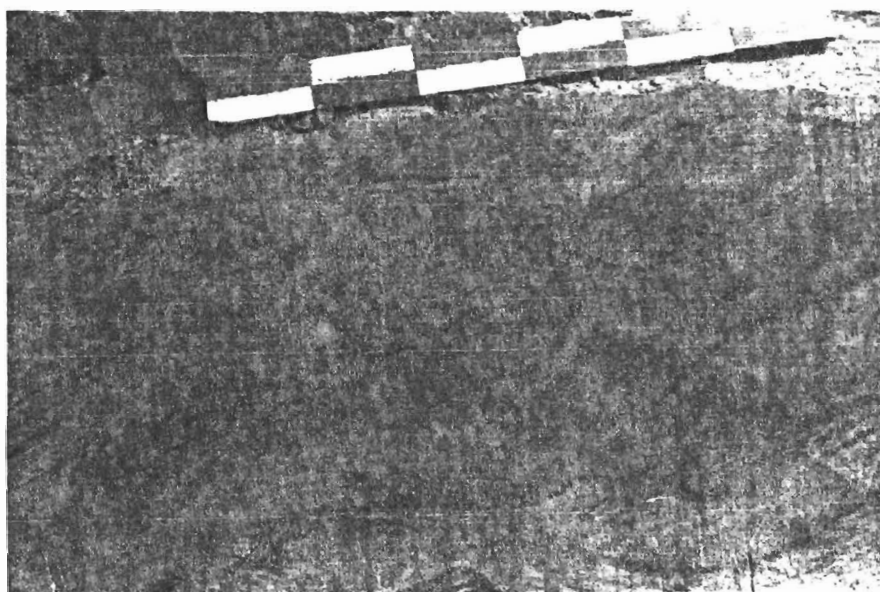
(A) type A climbing ripples.

(B) type B climbing ripples. Scale in 5cm graduations.

A



B



However, not all parts of the inclusion are in their pristine state. Figures 2.10 (a) and 2.11 (a), (b) and (c) are taken of the inclusions in figures 2.8 (a) and (b). Subhorizontal fractures clearly dissect subvertical layering, which may have been horizontal initially. Mixing of the sediments comprising the sand blocks can be readily seen in figures 2.11 (a), (b) and (c). In this instance, the deformed material is found to overlie the undeformed sediment. The boundary between the undeformed and deformed layers in the inclusion in figure 2.8 (a) is marked by the presence of a Dmm layer which can be seen to deform the subjacent Sr facies into an Sm facies (fig. 2.11 (d)).

The inclusion depicted in figure 2.8 (c), has apparently been tilted almost upright. Layering in figures 2.11 (e) and (f) is at a subvertical orientation; subhorizontal faults (fig. 2.10 (b)) cut this layering. Contact relationships with Dmm are also subvertical (figs. 2.12 (a), (b)).

The most intensely deformed inclusions show total destruction of primary structures resulting in Sm or, more commonly, Sd. These facies are prevalent in those inclusions depicted in figures 2.8 (d), (e) and (f).

As is shown in figure 2.10, brittle fracture is quite common in the sand facies. Both normal and reverse faults are present, and displacements rarely exceed 10.0cm. With increasing macro-scale deformation, brittle failure of the sediments decreases; with increasing deformation, inclusions behave in a more ductile fashion.

Figure 2.10. Brittle fracture in sand inclusions.

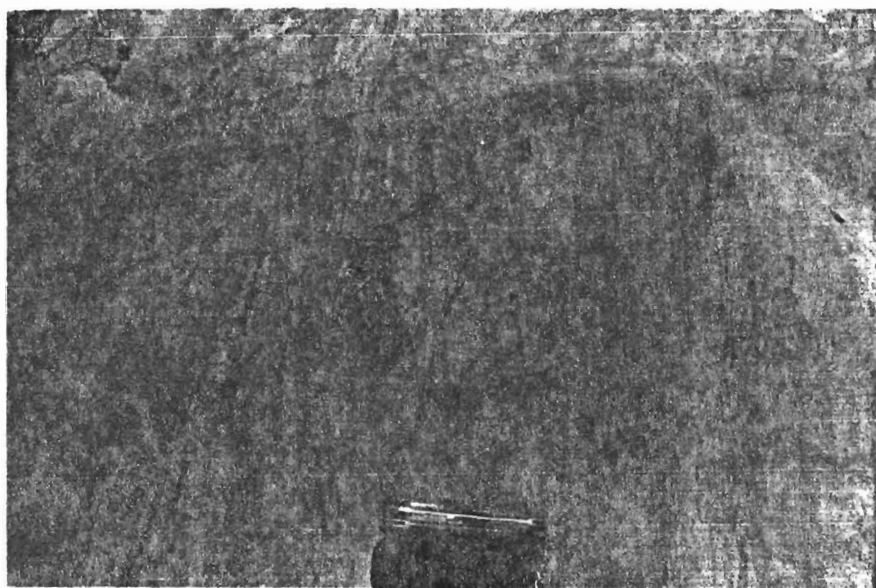
(A) Close-up of inclusion in figure 2.8 (B). Subhorizontal fractures dissect subvertical layering. Note shear zone in upper part of photograph. Knife for scale.

(B) Close-up of inclusion in figure 2.8 (C). Subhorizontal faults cutting a diamict layer that separates two distinct sand facies.

(C) and (D) Close-up of inclusion in figure 2.8 (A). Normal and reverse faults deforming Sr lithofacies. In (C) the intensity of deformation is clearly shown in the fault zone on the right side of photograph.

(E) Small scale faults within intercalated sands and diamict. Normal and reverse faults shown.

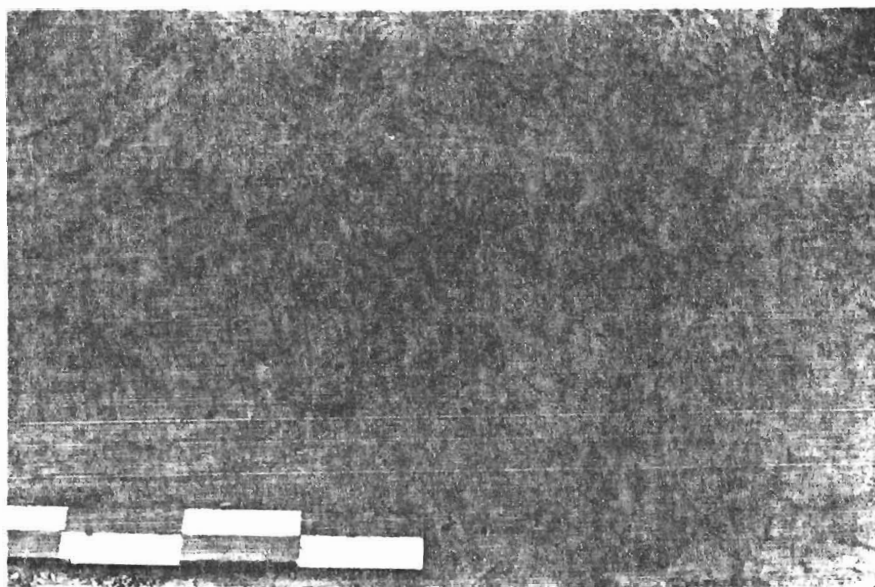
A



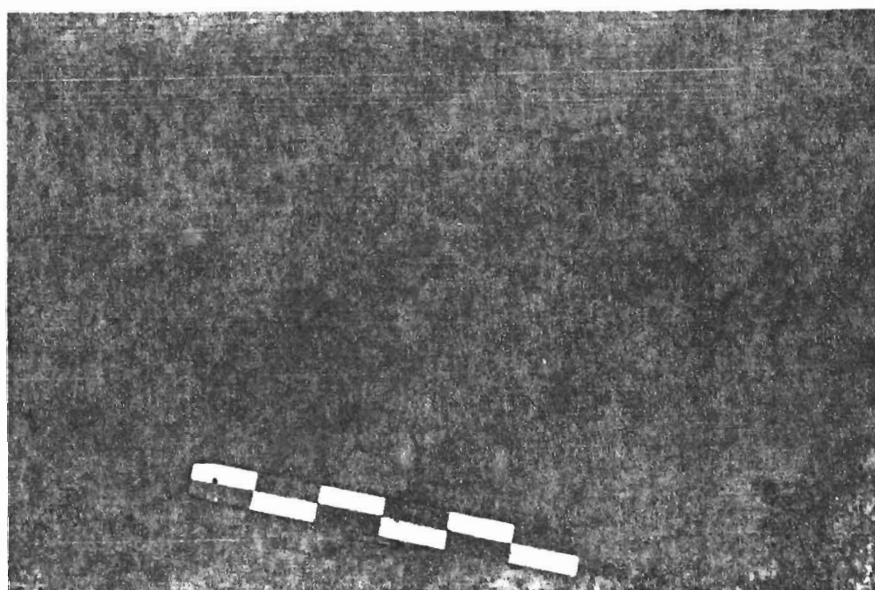
B



C



D



E

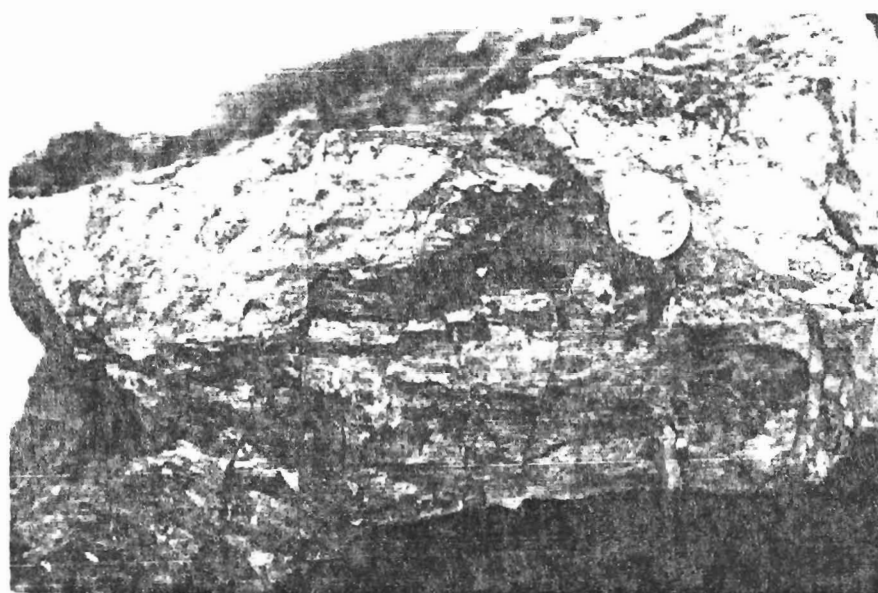


Figure 2.11. Deformation within sand inclusions.

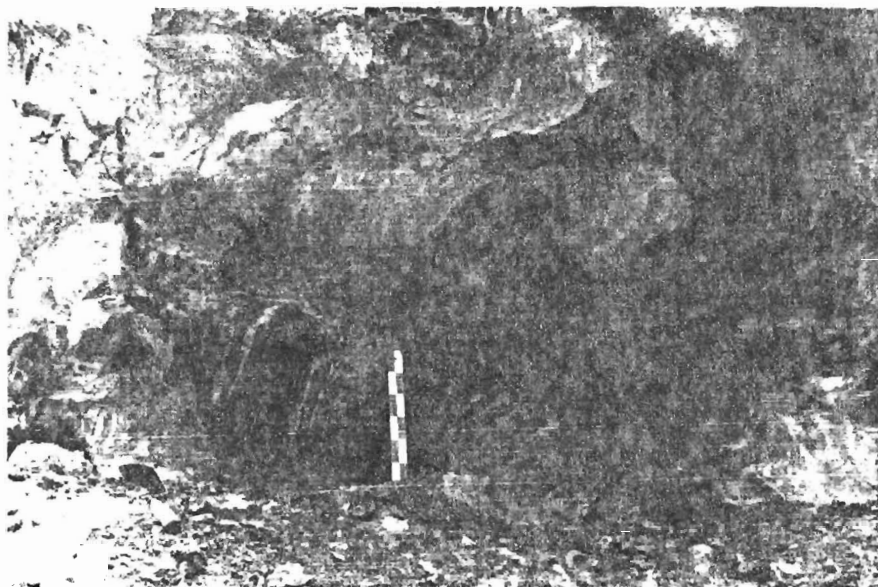
(A) and (B). Mixing and plastic deformation of sands. Photographs show the upper part of inclusion in figure 2.8 (A).

(C) Small scale shears within sands. Note that these shears are at different orientations and cross-cut other shears. Area of photograph is approximately 5x7cm.

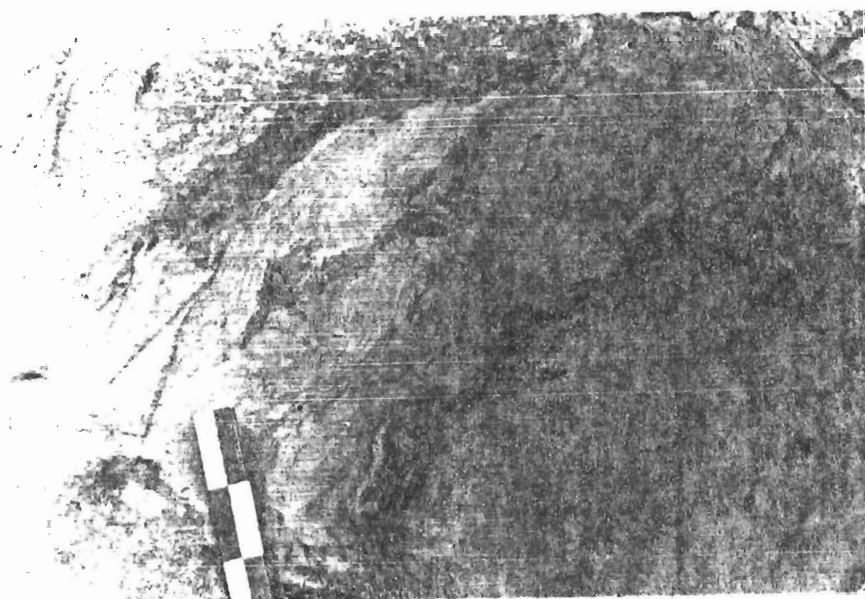
(D) Contact between the upper deformed part and lower undeformed part of inclusion in figure 2.8 (A). Dmm layer separates the upper and lower sections. The lower Sr facies experiences increasing deformation as the Dmm layer is approached until it becomes massive.

(E) and (F) Close-up of inclusion in figure 2.8 (C). Note subvertical layering, diamict inclusions, and rip-up clast breccia. Further, deformation is shown to increase towards diamict inclusion deforming sand into an Sm lithofacies. Knives for scale.

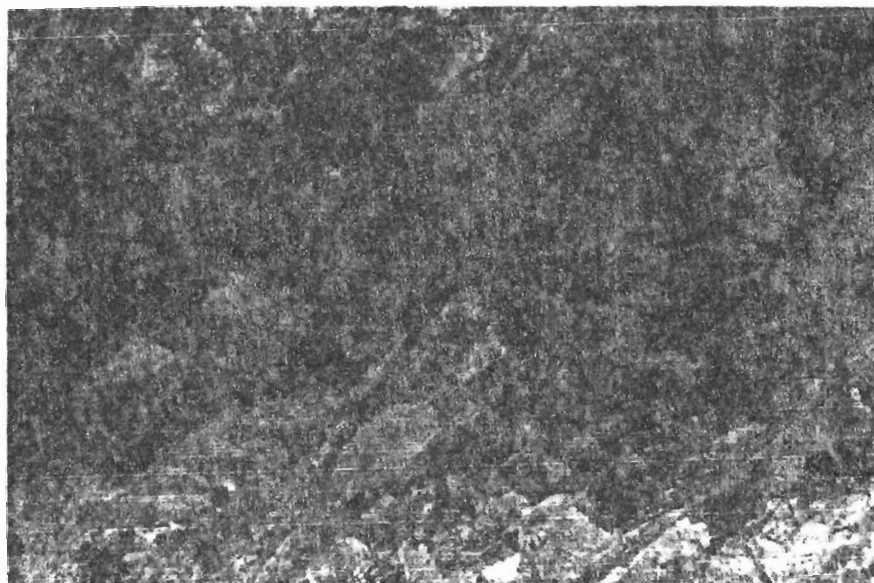
A



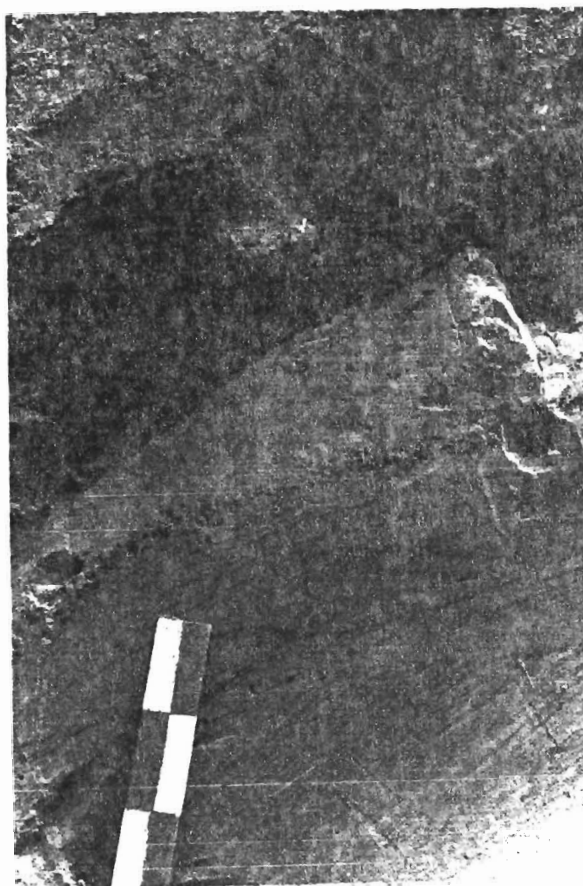
B



C



D



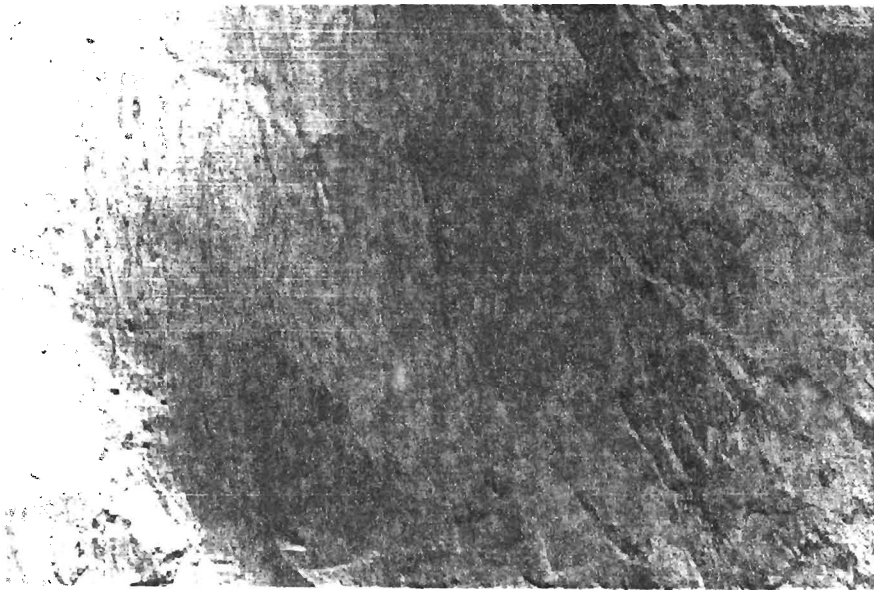
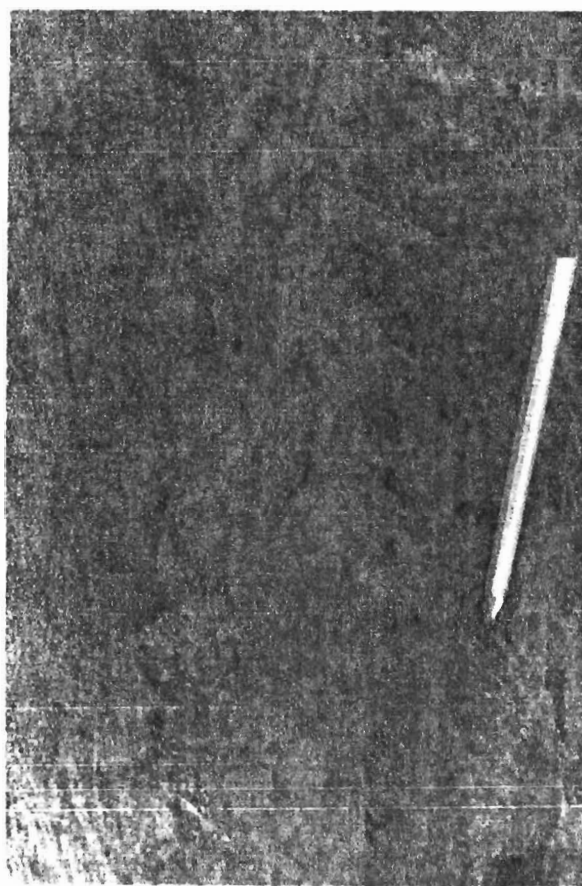
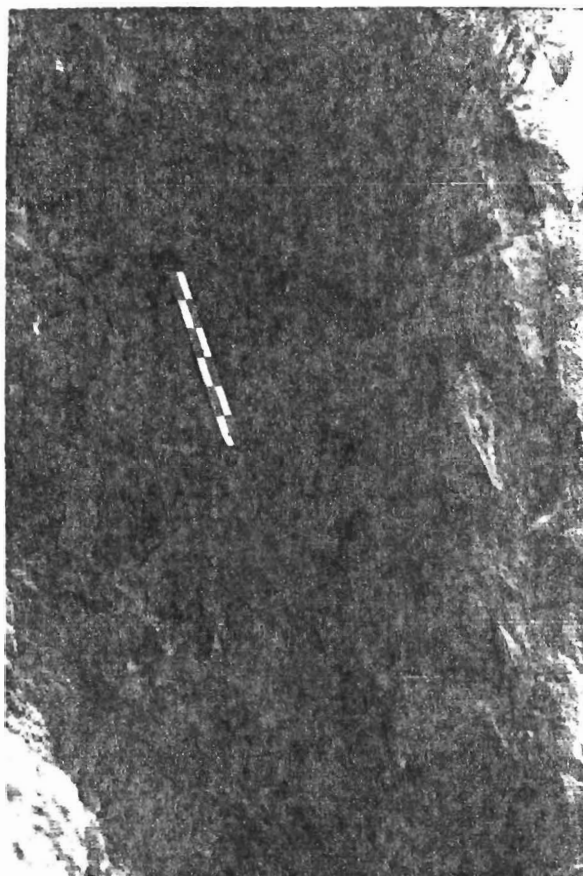
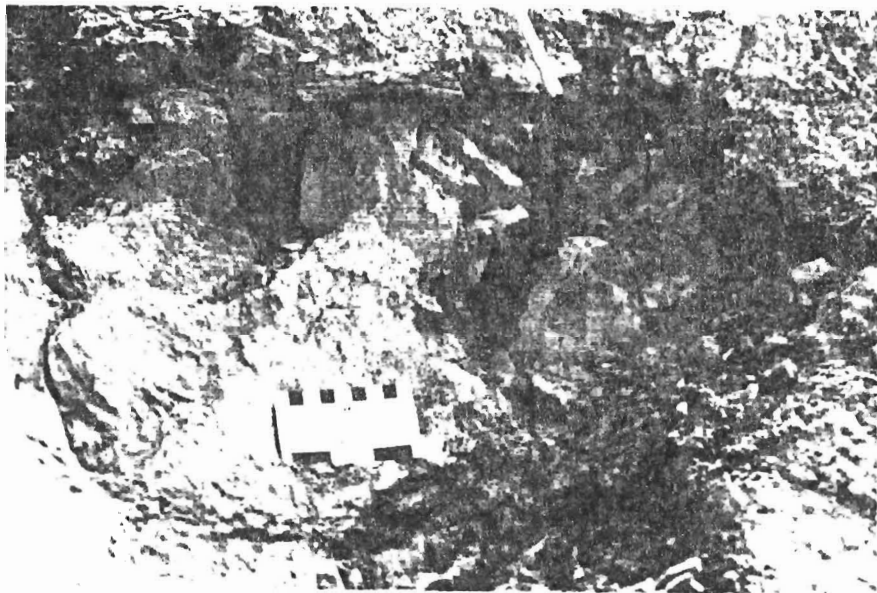
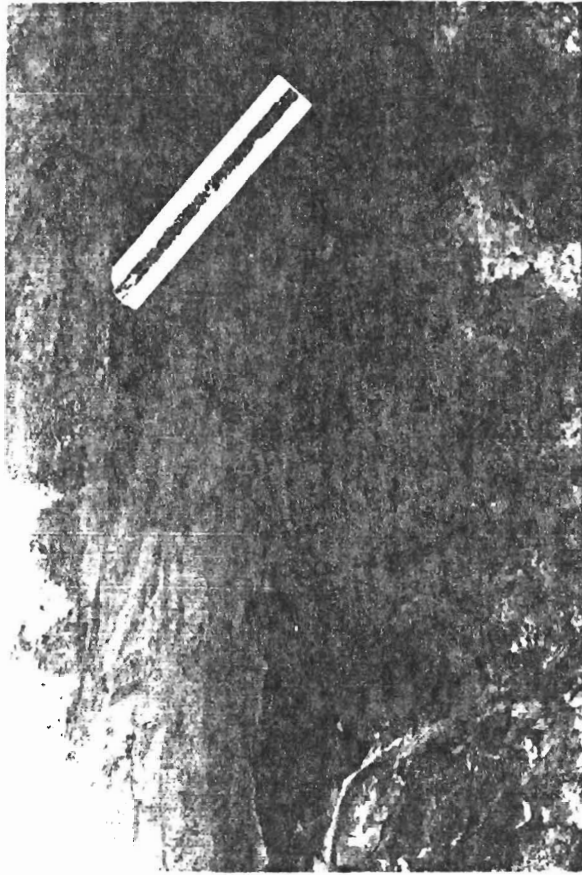


Figure 2.12. Sand diamict contact relationships.

(A), (B) and (C). Subvertical contacts. Penetrative deformation increases towards contact. In (B) and (C) boudinage can be observed.

(D), (E) and (F). Subhorizontal contacts. In (D) a diamict injection is shown, probably resultant from fracture exploitation. Faulting and deformation is readily apparent in (E). Note the abrupt truncation of the bedding in (F) by the overlying Dmm in contrast to the more plastically deformed contacts illustrated in (A), (B), (C) and (E).

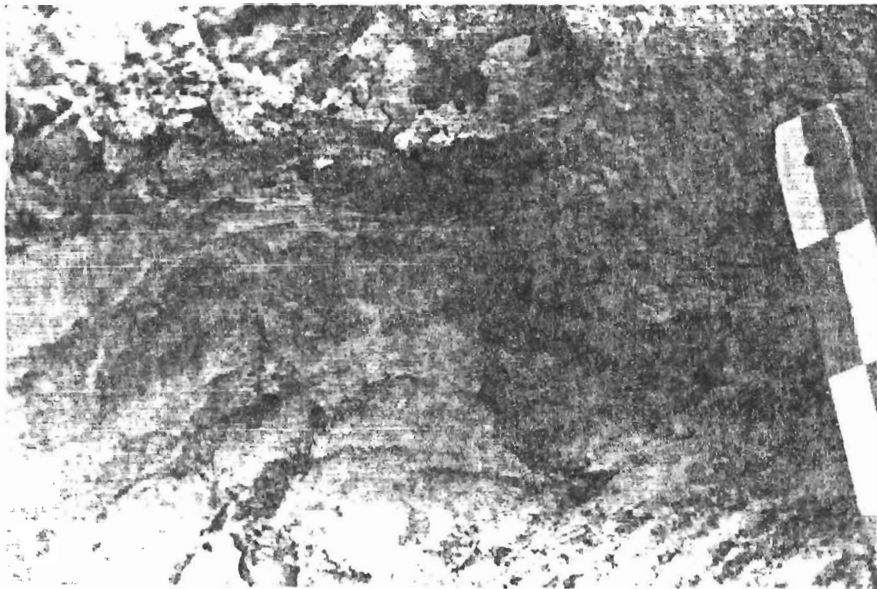




E



F



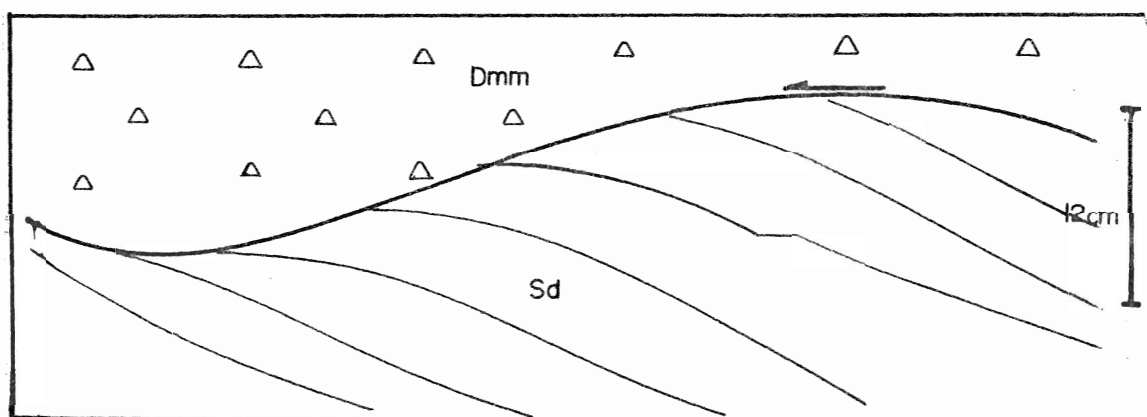


Figure 2.13. Field sketch of sheared Dmm/Sd contact. Note the apparent sinistral movement indicated by the deformed layering.

Contact relationships between the inclusions and overlying Dmm are varied (fig. 2.12). Planar to curvilinear contacts are most common although undulatory and irregular forms may occur. In the majority of cases, contacts are sharp, truncate subjacent layering, and show increasing penetrative deformation towards the contact. Geometries of the underlying layering sometimes indicate a simple shear with an overriding motion towards the southwest. In these cases, the sands have been deformed into Sd lithofacies.

Interlayering of the contacts can occur (fig. 2.14). Due to the apparent attenuation of the sand layers, it is suspected that this layering is of post-depositional origin produce by a similar mechanism that formed the sheared contacts. Elaboration of this idea will follow in the next chapter. ✓

Lineations have been produced at the contact between the sand inclusion and the overlying Dmm, within the plane of the contact. These lineations are similar to those documented by Ehlers and Stephen (1979) and Shaw (1982). Unlike Ehlers and Stephen (1979) who have subdivided these features, the author feels that the term lineation as defined by Hobbs et.al. (1977) is more appropriate since it encompasses all linear elements encountered and avoids any genetic inference. Azimuth and plunge orientations were measured and plotted on an equal-area projection (fig. 2.16). Azimuth orientations trend northeast-

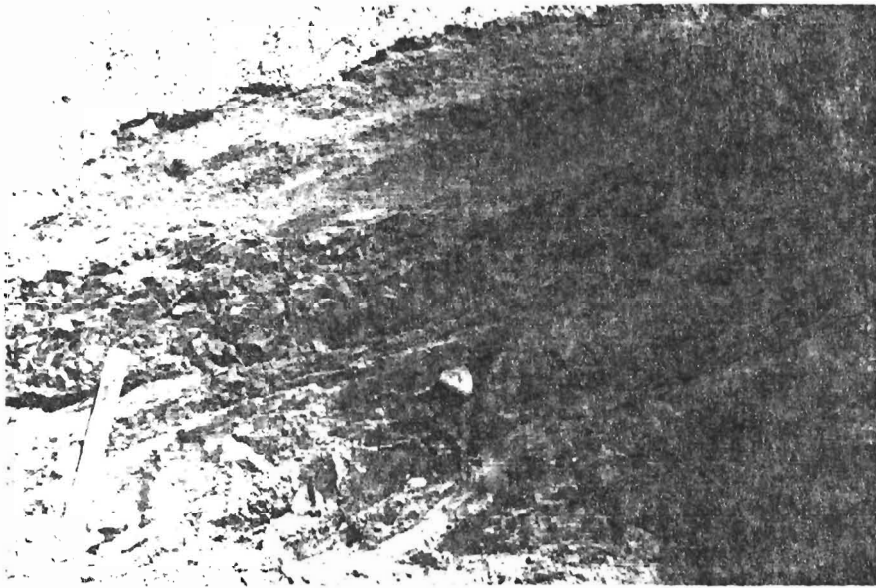
Figure 2.14. Interlayered sand/diamict contacts.

A) Note apparent attenuation of Sm layers and disaggregation from the lower inclusion.

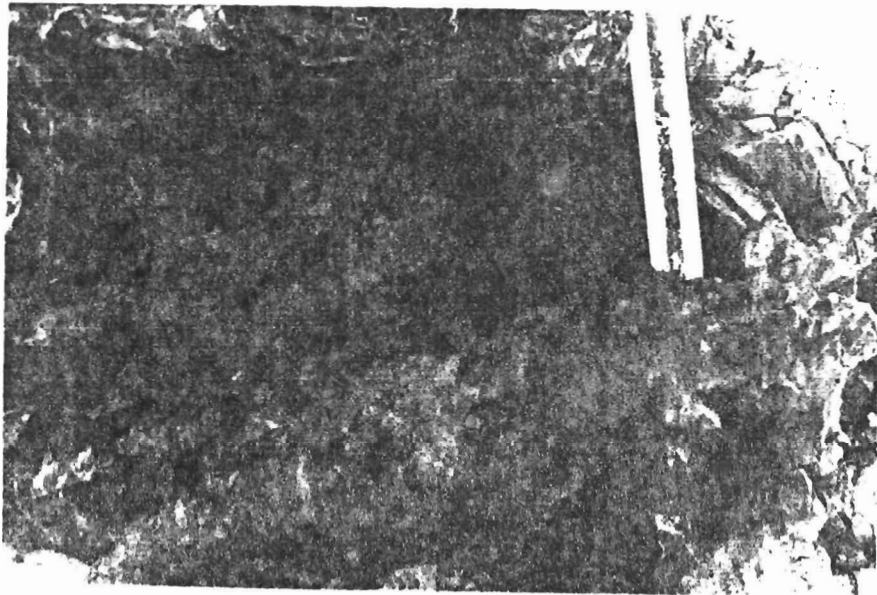
B) Note incorporation of sands within Dmm and Dmm inclusions in sand.

C) Laboratory photograph of impregnated sample. Small scale disaggregation can be seen along with attenuation. Brittle fracture is attributed to sample preparation.

A



B



C

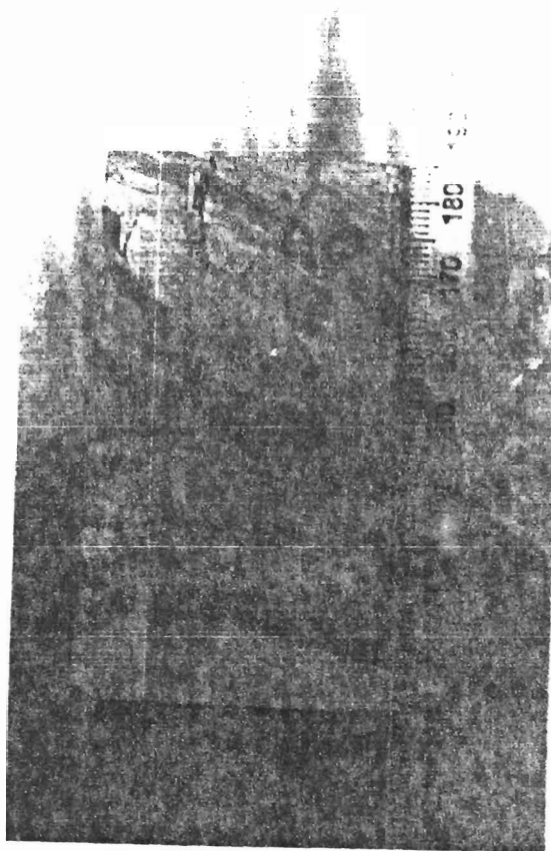
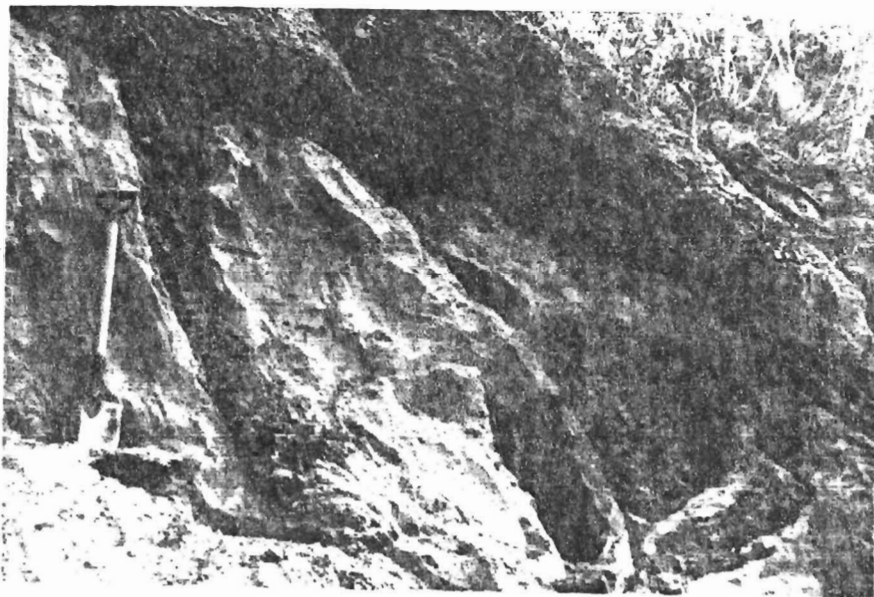


Figure 2.15. Diamict injections. Note destruction of primary structures in sands.

A



B



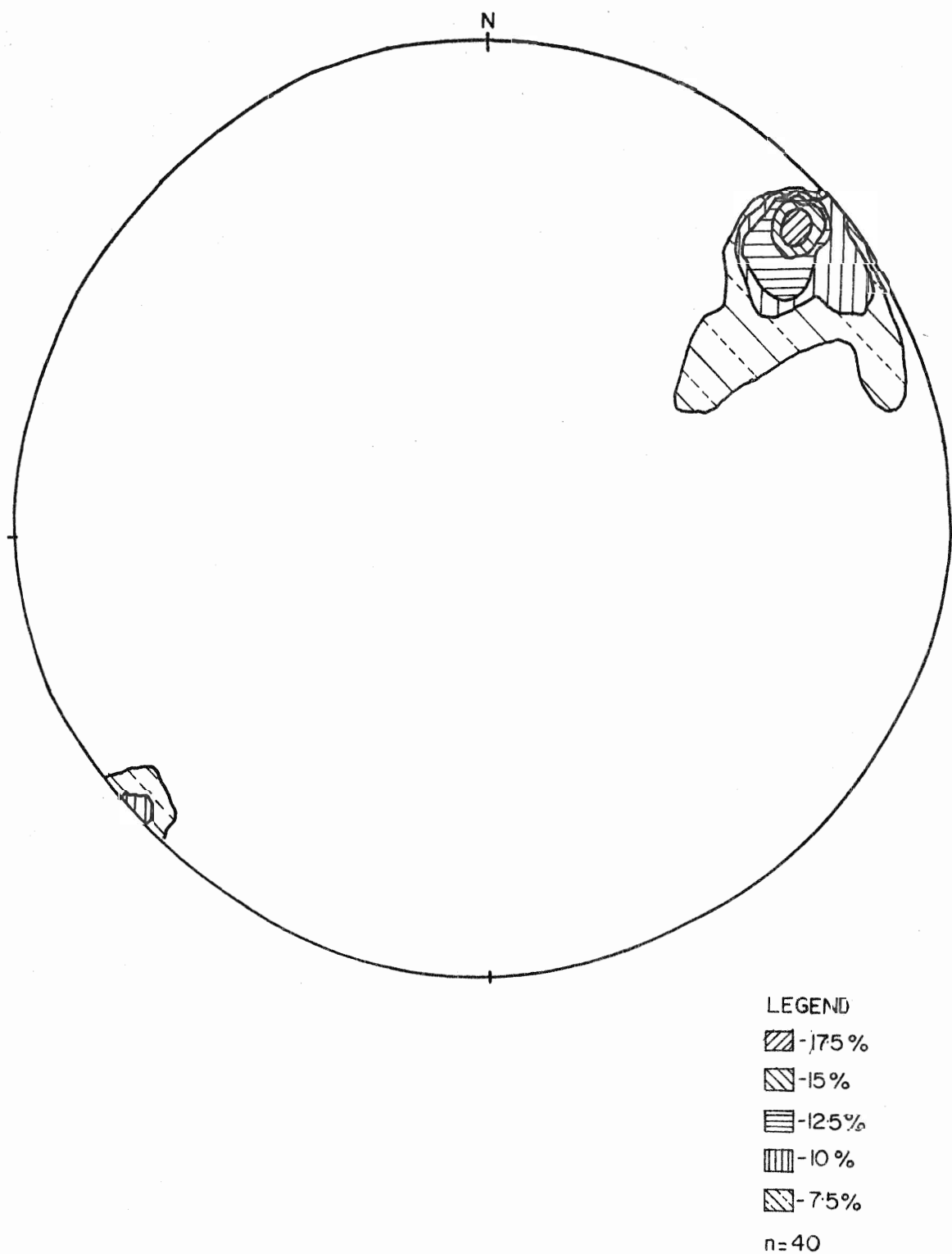


Figure 2.16. Equal-area projection of contact lineations.

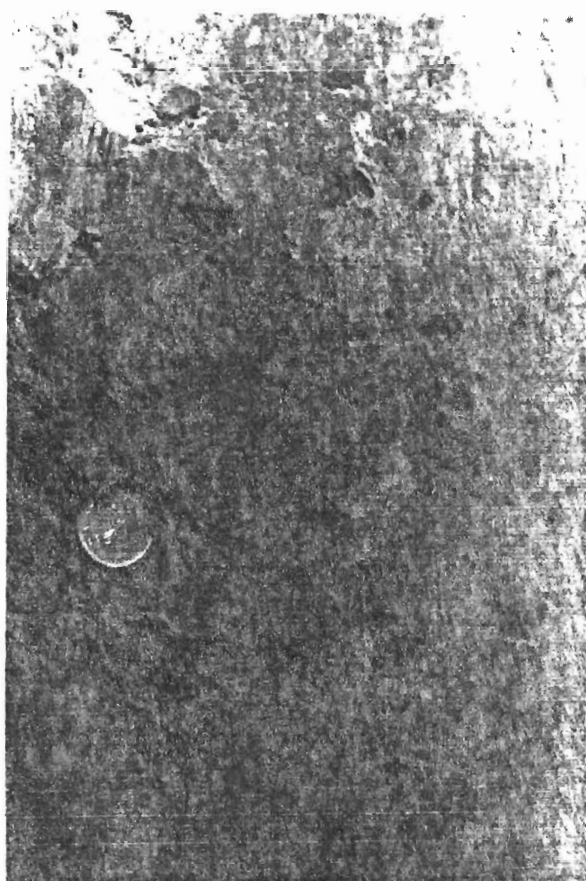
southwest with a preferred plunge of northeast.

Injections of the overlying Dmm into the inclusions are commonplace (fig. 2.8, 2.11 (d), 2.15). As is shown in these figures, these injections take many forms and are probable due to fracture exploitation. Primary structures are usually destroyed in the process. ✓

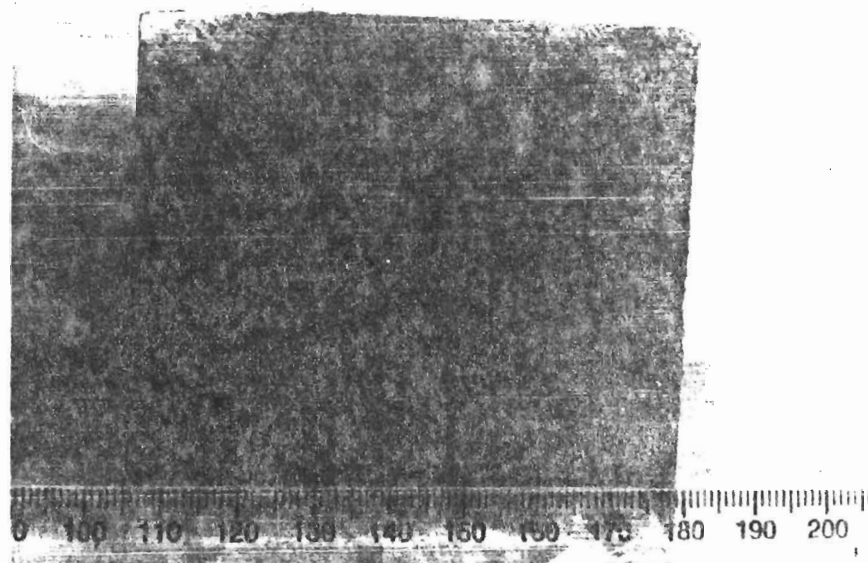
Finally rip-up clast breccias were encountered in localized areas of certain sand inclusions. They were usually associated with Sm (fig. 2.17) but were found to occur occasionally with Sr. In the latter instance, the clasts did not deform the bedforms, thus appearing to be a syn-depositional feature, as opposed to "dropping in" from suspension.

Figure 2.17. Rip-up clast breccia. Small-scale faulting also present in (A).

A



B



Unit B

2.2.2

Stratigraphic unit B is a relatively minor unit which is only exposed along the flanks of the exposure (fig. 2.1). It is composed of three discontinuous lithofacies: Dmm, Dmm(r)/Fmd and Sm. The thickness of the unit does not exceed 4.5m; individual lithofacies are rarely thicker than 1.0m. Dmm, Dmm(r)/Fmd, and Sm all interdigitate without any apparent cyclicity. The only general statement with regards to depositional patterns is that Dmm is dominant towards the ridge whereas with increasing distance along the flanks, intercalations of Dmm(r)/Fmd and Sm increase with Dmm(r)/Fmd forming the uppermost layer at the extremities of the exposure.

The contact between units A and B was very difficult to locate due to present-day slope failure. Due to the similarities of the Dmm lithofacies of the two respective units, the transition was impossible to distinguish. Therefore the first incursion of Dmm(r)/Fmd was used as the base of unit B. Where exposed, this contact was sharp and unconformable with a complete lack of deformation.

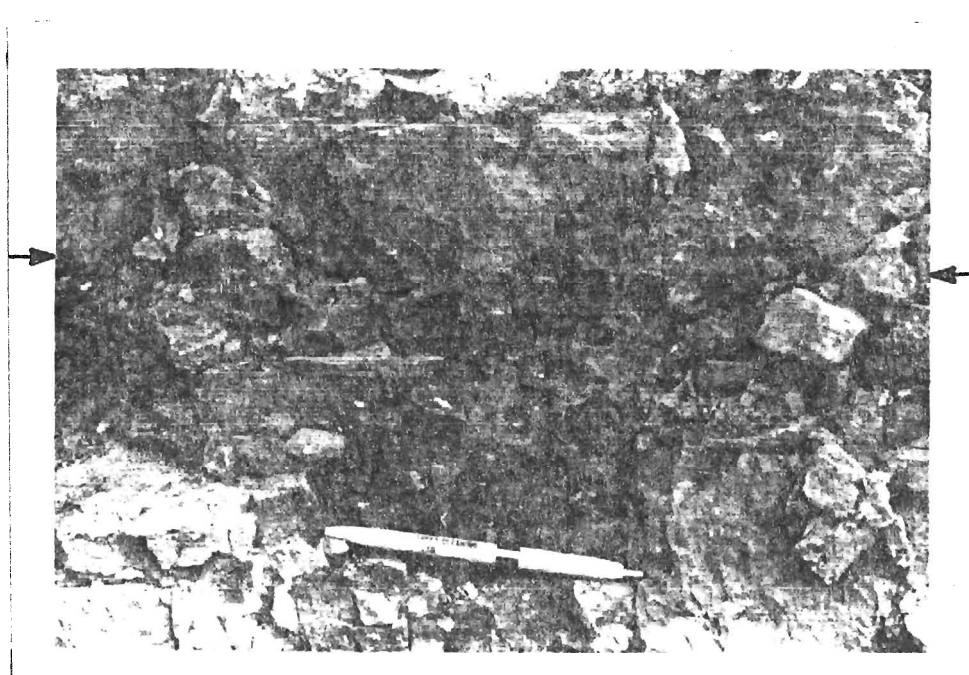
Dmm is virtually identical to its counterpart in unit A (fig. 2.18). However there are distinct differences to be made note of. Sieve data (fig. 2.4) indicates that

Figure 2.18. Dmm and Dmm(r)/Fmd. In both photographs Dmm is overlying Dmm(r)/Fmd. Contact illustrated by arrows. Note the amount of silt clasts within Dmm(r)/Fmd is far greater than Dmm, and lower clast contents.

A



B



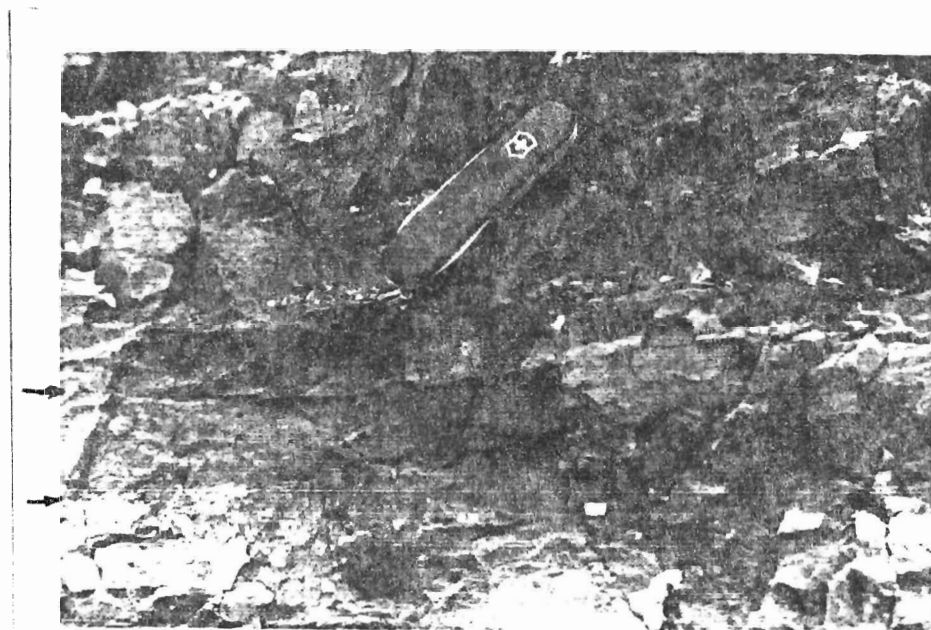
it is a finer-grained diamict than that of unit A. Clast contents appear to be lower, along with a drop in the number of striated clasts and an increasing degree of roundness being observed. Silt clasts are abundant throughout. Dmm is more friable in unit B and is less dense; both characteristics are attributed to soil forming processes. Iron-staining, commonly found along fissures, is quite common as well as the complete breakdown of clasts, notably exotic lithologies such as granites. These features, and others, have been mentioned by Madgett and Catt (1977) and Eyles and Sladen (1981) both of whom have documented the effects of pedogenic processes to depths of 8.0m in tills in northern England.

The other diamict lithofacies, Dmm (r)/Fmd (fig. 2.19), is a fine grained sediment (fig. 2.4). The amount and size of clasts is far less than Dmm; they tend to be ≤ 1.0 cm (with a few exceptions), well rounded and non-striated. Silt clasts are quite common. Like Dmm, certain pedogenic features are present but not to the degree to which they effect the former diamict. This may be due to the very fine grained nature of Dmm(r)/Fmd: the decrease in permeability due to high silt/clay contents will hinder the movement of groundwater, thus the amount of pedogenic alteration will be less. ✓

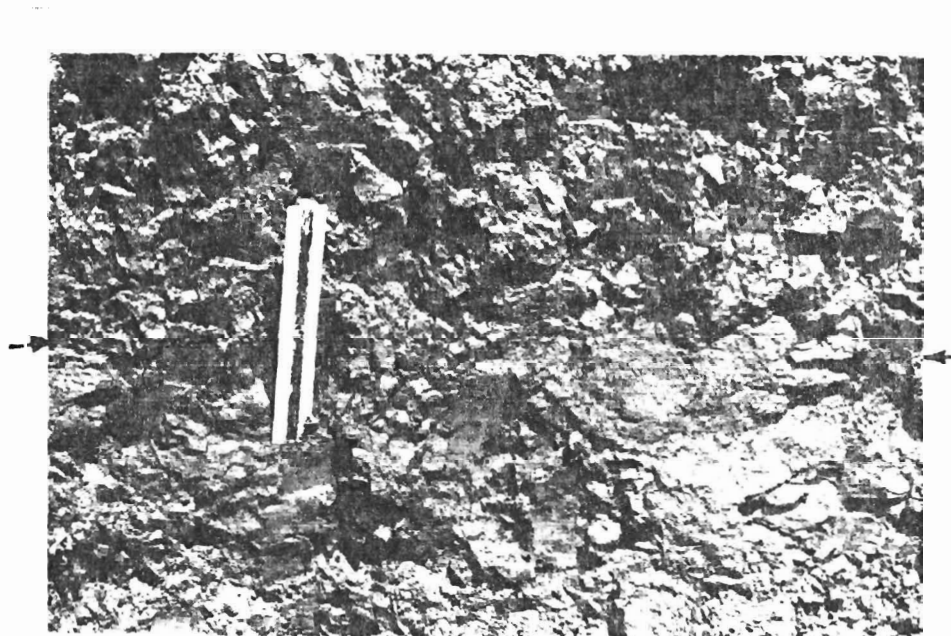
The reason for assigning two codes to this sediment is related to its change in physical properties with stratigraphic height. At lower stratigraphic levels, the

Figure 2.19. Rafting of Dmm in Dmm(r)/Fmd. Contacts indicated by arrows.

A



B



aforementioned characteristics are well developed.

With increasing height the abundance of both silt and rock clasts decrease to practically nil and pedogenic alteration increases. Alteration is an obvious fact, but the drop in clast content of both types may reflect a change in both sediment and depositional process. The sediment becomes less diamictic in appearance and more like a massive, fine grained sediment. The presence of the occasional clast may be due to a "dropping-in" through a water column.

Rafts of either diamict may be hosted by the other. Figure 2.19 (a) illustrates a case in point where a raft of Dmm is present within Dmm(r)/Fmd. These rafts tend to be small, usually $< 1.0\text{m}$ across and $< 20.0\text{cm}$ in thickness. Most are lens shaped. The amounts of Dmm rafting within Dmm(r)/Fmd decrease with stratigraphic height lending more credence to the idea that depositional processes were changing over time.

Sm is a rather mundane lithofacies which occurs in a variety of forms. Figure 2.20 cites two examples, where it occurs as a large lens-shaped layer and as discontinuous blocks. This lithofacies is far different than the intra-formation inclusions in unit A. Notably the variety of deformation styles are absent: this sand facies appears totally devoid of deformation of any type. Contact relationships, although sharp and unconformable with the diamicts, do not show any evidence of shear or alteration

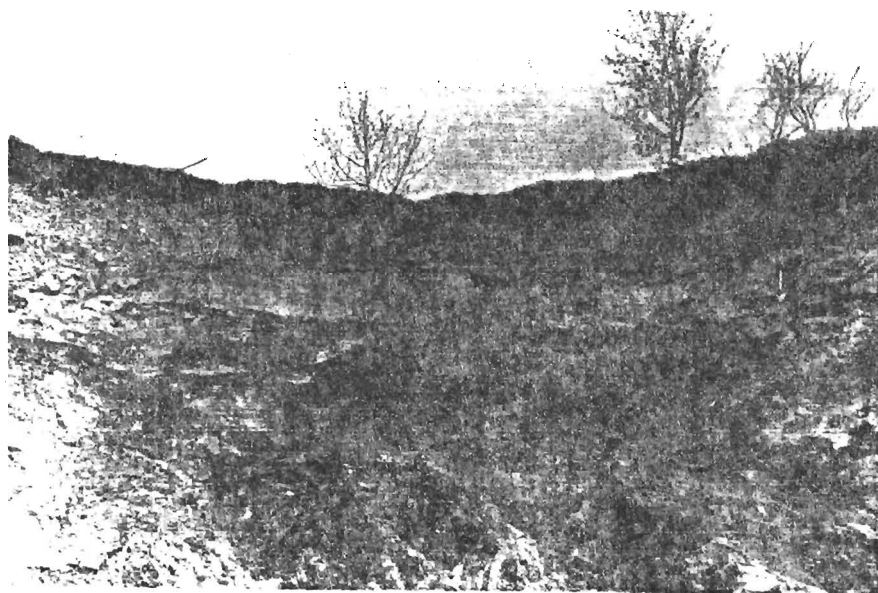
Figure 2.20. Sm lithofacies.

(A) large, lens-shaped layer

(B) discontinuous inclusions.

(C) and (D) contacts with Emm(r)/Fmd.
Note complete lack of glaciotectionic
deformation. Contact in (D) shown by
arrow.

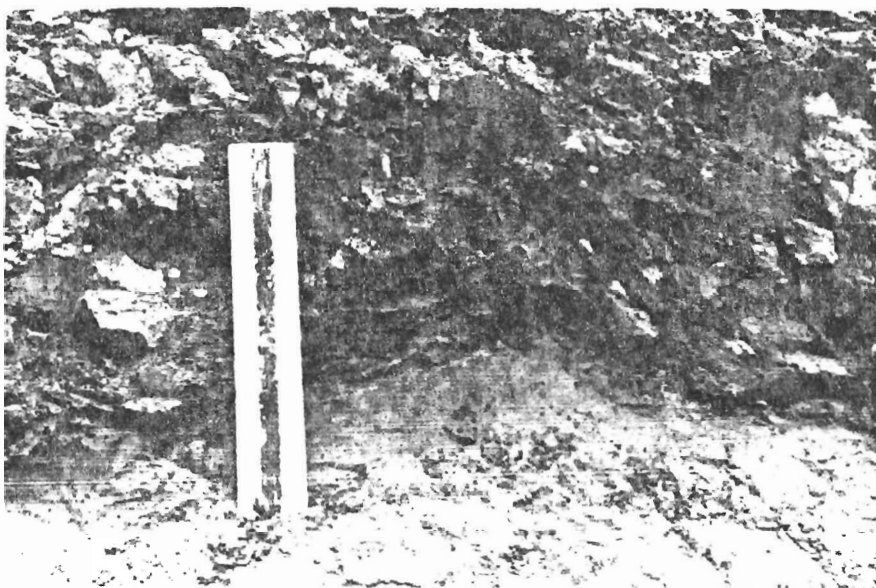
A



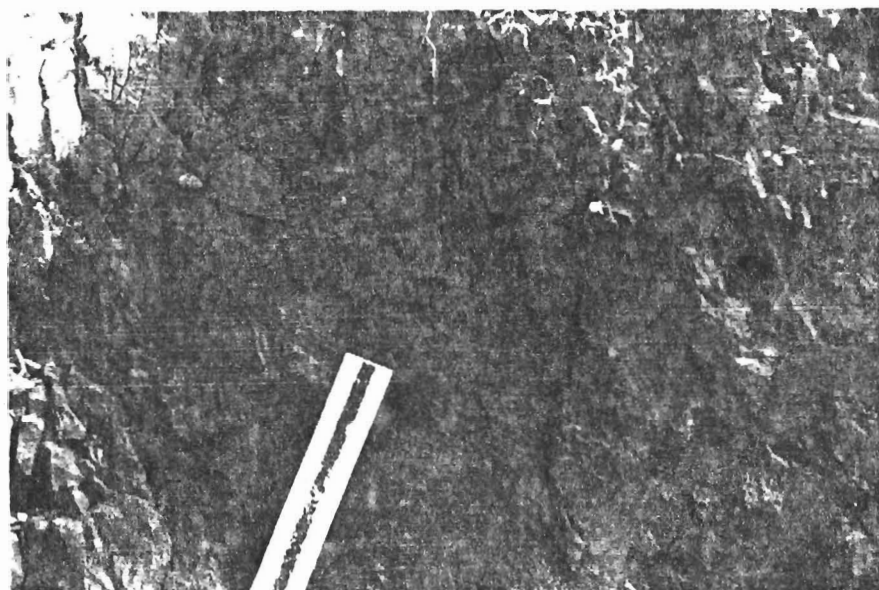
B



C



D



that characterize the sand/diamict contacts of unit A (fig. 2.20). Thus the geometry of the Sm must be related to depositional as opposed to post-depositional mechanisms.

Further differences with unit A inclusions are found in grain size analysis (fig. 2.4). Grain size tends to be highly variable. This may, once again, be a function of pedogenesis. Due to the high permeability of a sand unit, percolation of groundwaters will be quite rapid. Thus fines may be flushed out of some units and concentrated in others which overlie a (more) impermeable unit. Staining is evident in some layers. This may also be a reason for the massive nature of this sediment. Intensive diagenesis of the sands from pedogenic processes may have resulted in the destruction of primary structures (c.f. Madgett and Catt, 1977; Eyles and Sladen, 1981).

As was stated with respect to Sm/diamict contacts, relationships between Dmm and $Dmm(r)/Fmd$ are also quite sharp. Figures 2.19 and 2.20 illustrate these features. There is a complete lack of glaciotectionic deformation at these contacts, if rafting is not included as deformation. Even if it is, observation of these figures clearly shows the dissimilarity between these and the sheared contacts of unit A.

All three lithofacies exhibit flow characteristics. Flow noses were observed occasionally throughout unit B

and, although flow directions were difficult to obtain, a general southwest transport direction is inferred. Figure 2.21 gives an example of a diamict flow.

2.2.3

Unit C.

Stratigraphic unit C is located in the centre of the exposure, coinciding with the maximum elevation of the ridge (fig. 2.1) and attains thicknesses of up to 14.0m. This unit pinches out rapidly as the flanks are approached. Contacts with units A and B were difficult to find due to slumping. However an unconformable relationship was observed with unit A. The contact with unit B was never identified precisely and may be transitional. Although owing to the differences in lithofacies composition of the two units, it is doubtful that this is the situation. A more probable cause may be a post-depositional modification by lower water levels. This aspect and its implications will be explored further in proceeding chapters.

Unit C is composed of a variety of sand lithofacies with subordinate diamict and gravel lithofacies. Sand facies present are Sd, Sr, Sm, St and Sh. Sd is, by far, the most dominant within unit C (fig. 2.22). It is very

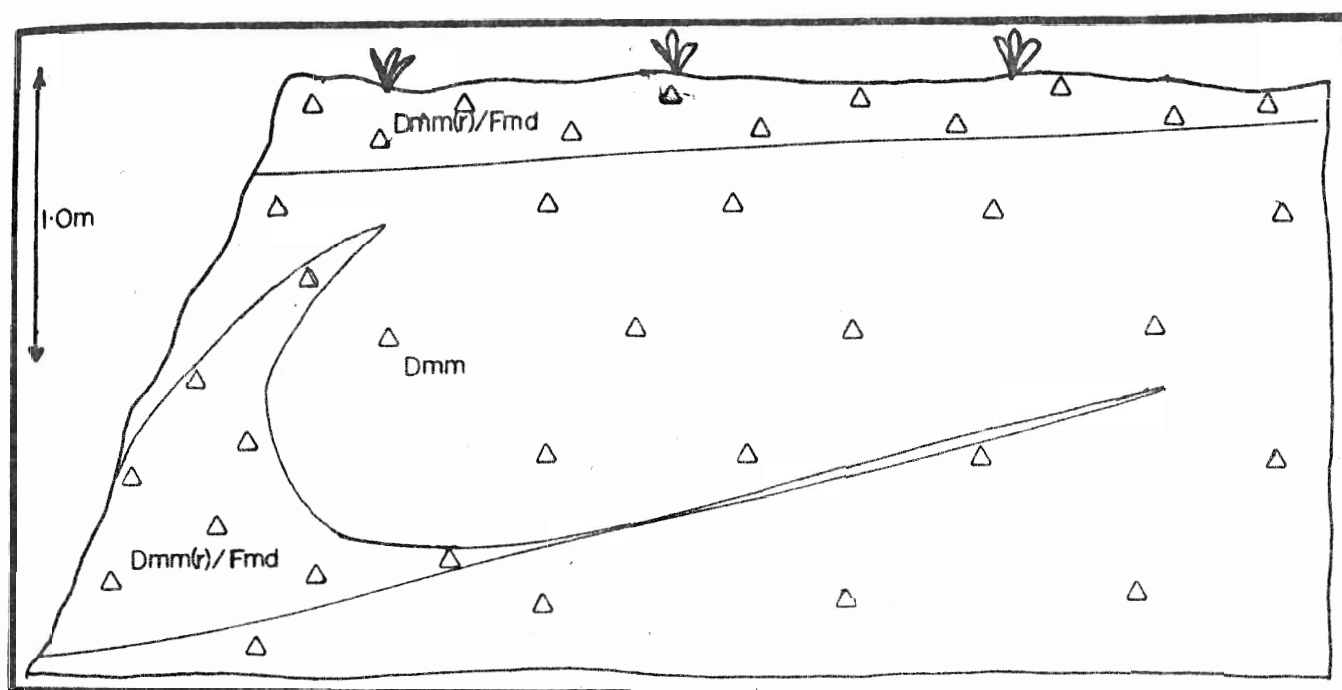


Figure 2.21. Flow structures in unit B. In this field sketch Dmm has flowed into a Dmm/Fmd deforming the Dmm(r)/Fmd into a push front.

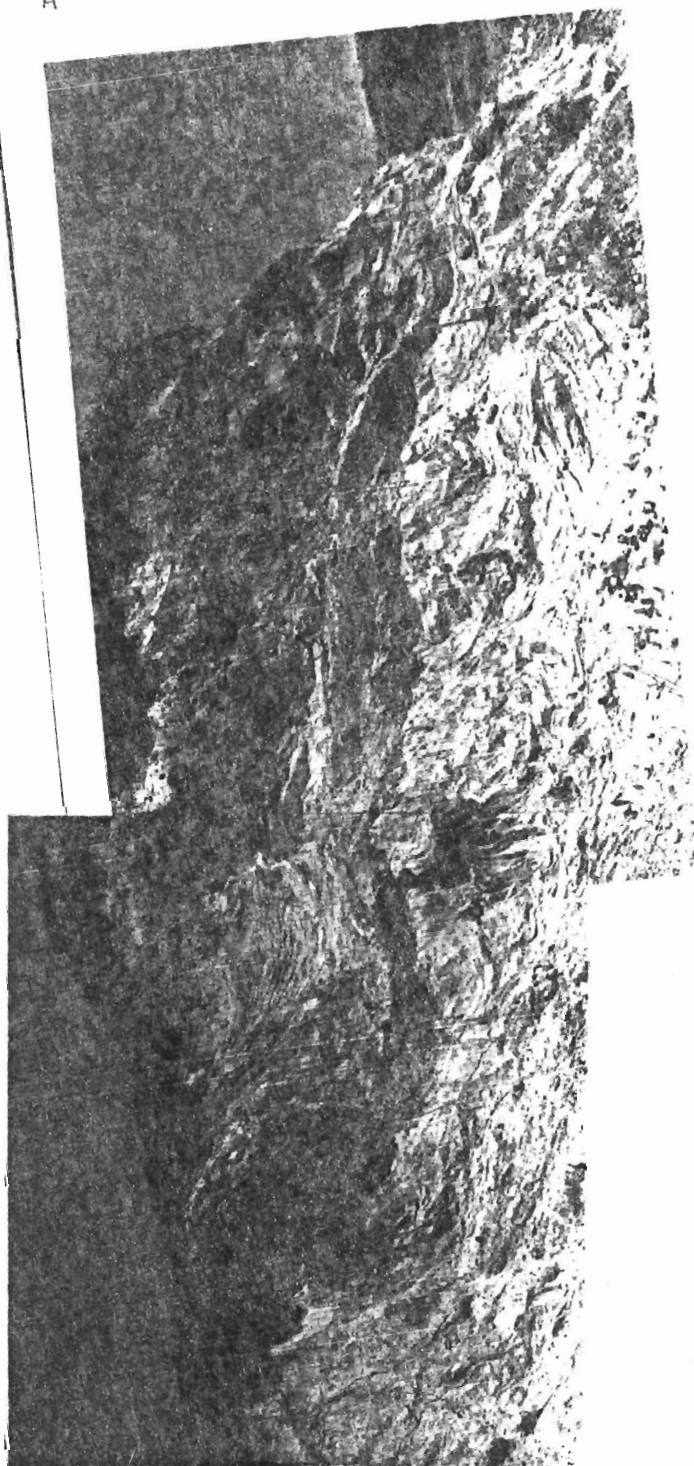
Figure 2.22. Soft-sediment deformation within unit C.

A) Mosaic of intercalated Dmm and Sd. Dmm is shown truncating water-escape structures in Sd.

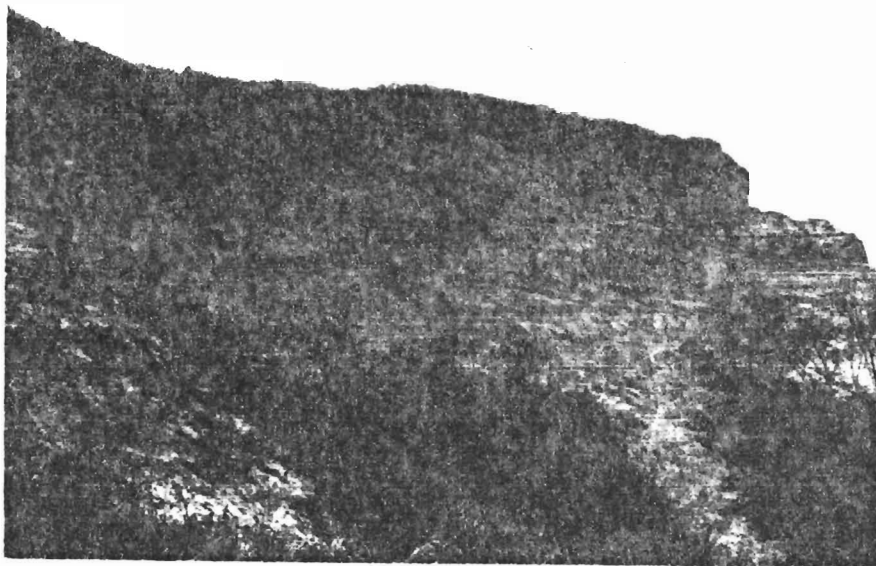
B) Various sand lithofacies with thin Dmm interbeds. Note how the units grade laterally from undeformed to highly deformed. Further, the scale and amount of dish-and-pillar structures should be observed.

C) Sd. Total destruction of primary structures and internal stratigraphy.

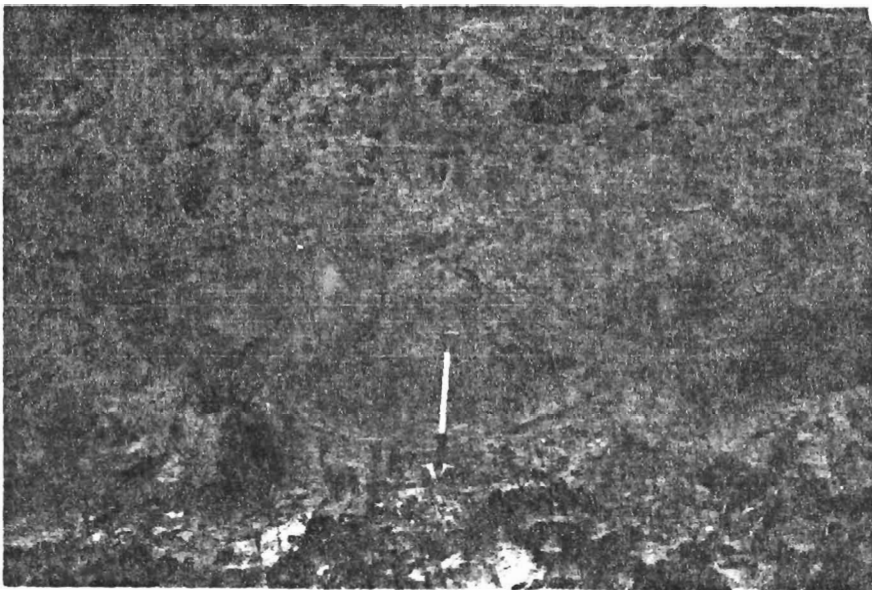
A



B



C



different from its counterpart in unit A. Sd is characterized by soft-sediment deformation in the form of dish and pillar structures, flame structures, loading (ball-and-pillow structures) and convolute laminations (fig. 2.23) characteristic of water release (Lowe and LoPicollo, 1974; Lowe, 1975). Brittle failure is also present (fig. 2.24). As is shown in figure 2.23, the size of some of these structures is unusually large. Most sizes of soft-sediment deformation structures cited from the literature (e.g. Lowe, 1975) are on the scale of centimetres. The dish and pillar structure depicted in figure 2.23 (a) was measured to have "dishes" in excess of 4.0m each and a pillar of over 2.0m in width. Most are quite a bit smaller, however structures of this type are commonly about 2.5m in total width, including the pillar. Dish and pillars and flame structures are often found in close spatial association: sequences of sediment over 10.0m in width and approximately 2.0m in thickness have been observed to be deformed by a series of dish and pillars with intervening flame structures.

Loading (ball-and-pillow structures) is common although not as prevalent as the dish and pillar and flame structures. This feature tends to be found in isolated places. By themselves, convolute laminations and brittle fracture are isolated occurrences, but when in association with other soft-sediment deformation

Figure 2.23. Soft-sediment deformation structures.

(A) and (B). Dish and pillar structures. In (A), note the presence of a gravel cut-and-fill channel in the upper part of the section and that it has not been effected by water-escape. In (B), the absence of primary structures should be observed.

(C) and (D). Flame structures. In (C), these are associated with dish-and-pillar structures.

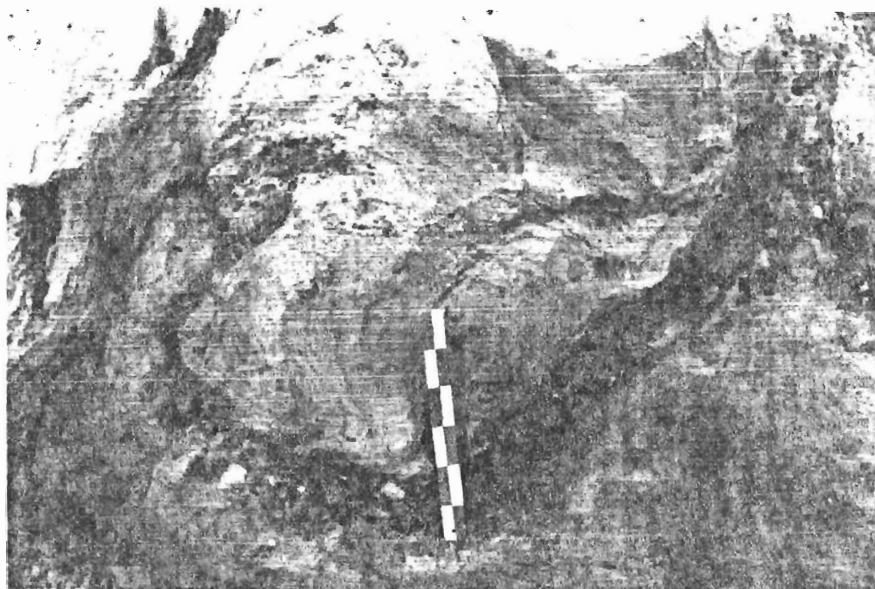
(E) and (F). Load structures.

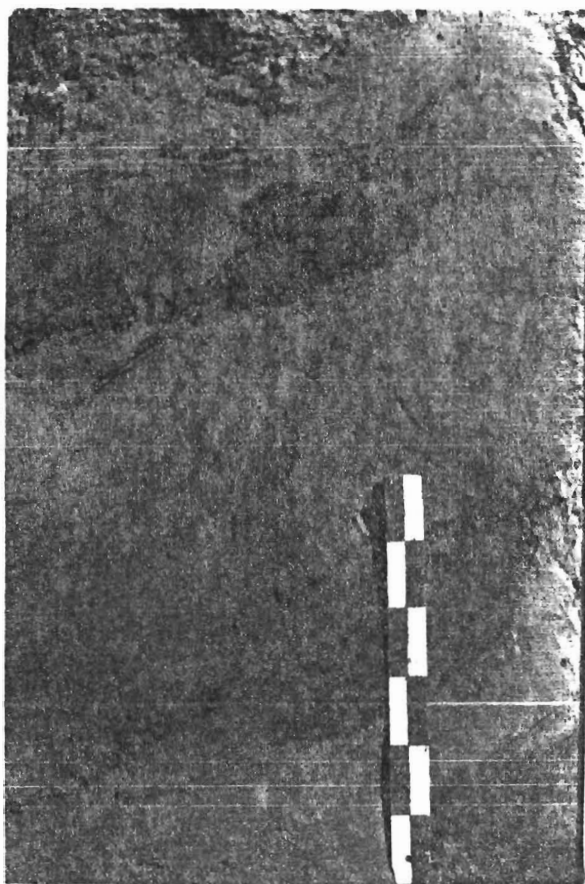
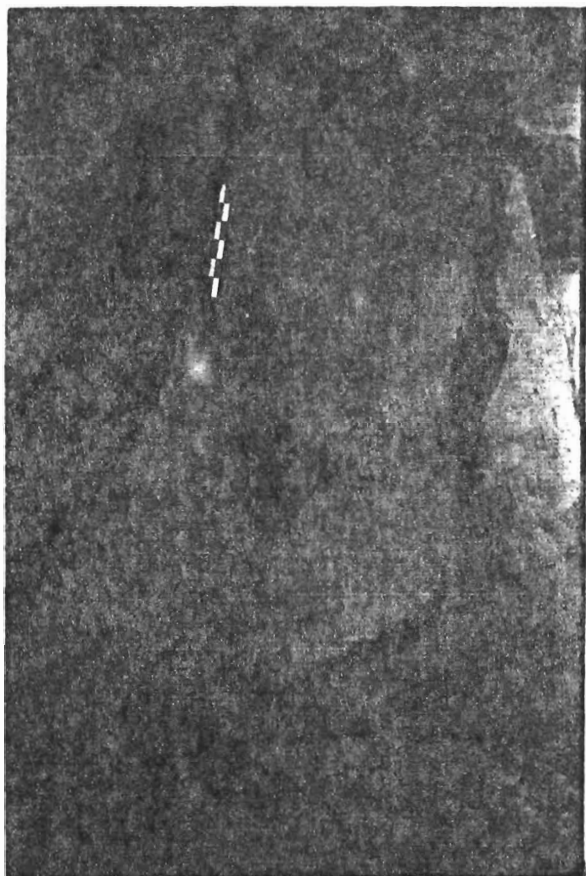
(G) Sr lithofacies (in this instance type B climbing ripples) grading into convolute laminations (Sd).

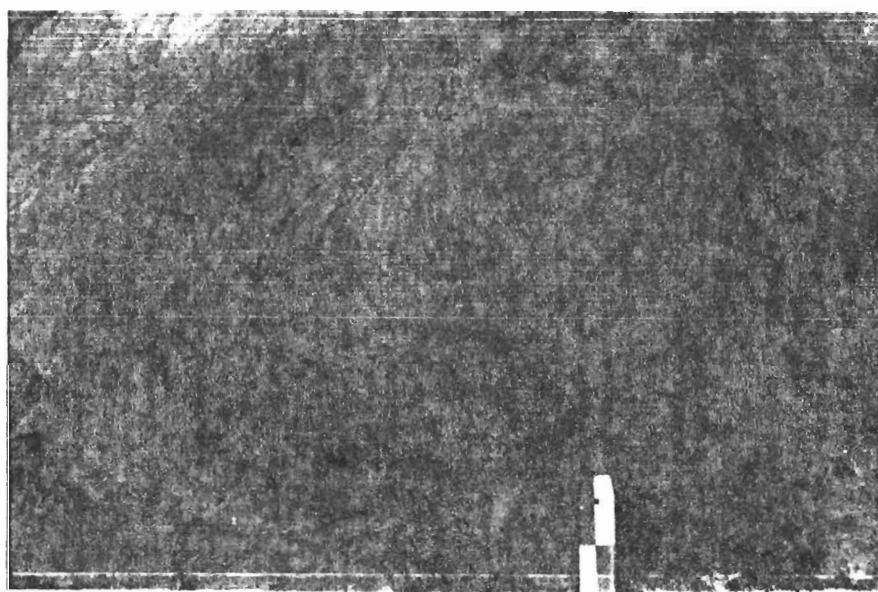
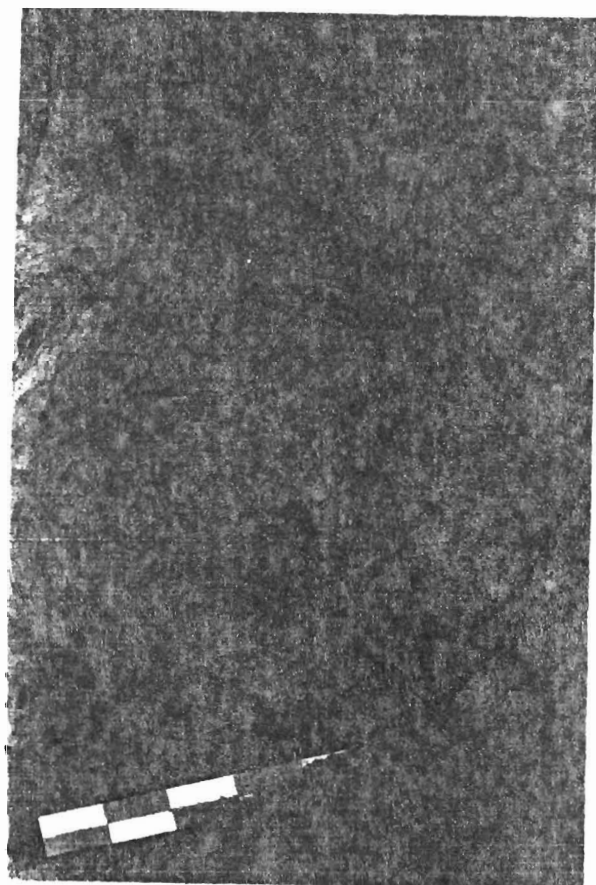
A



B







G

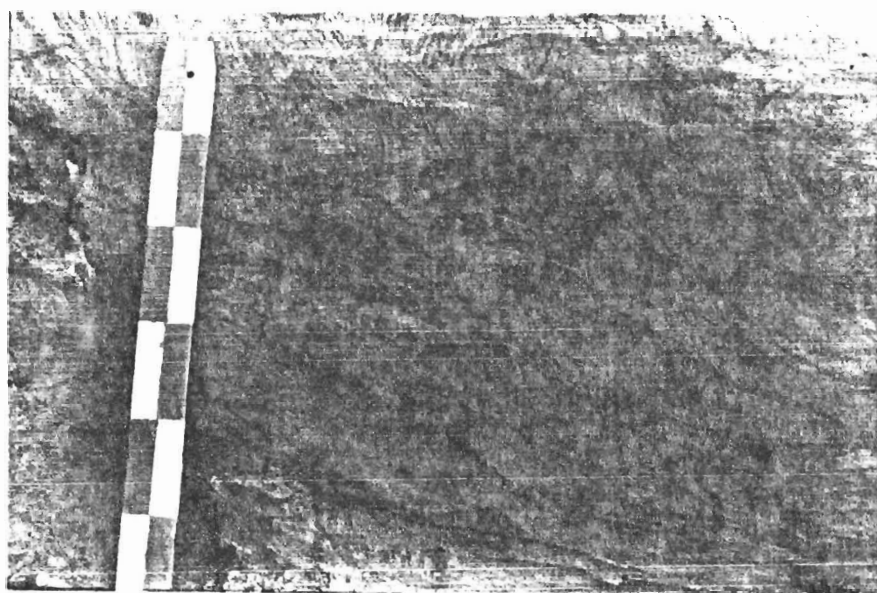
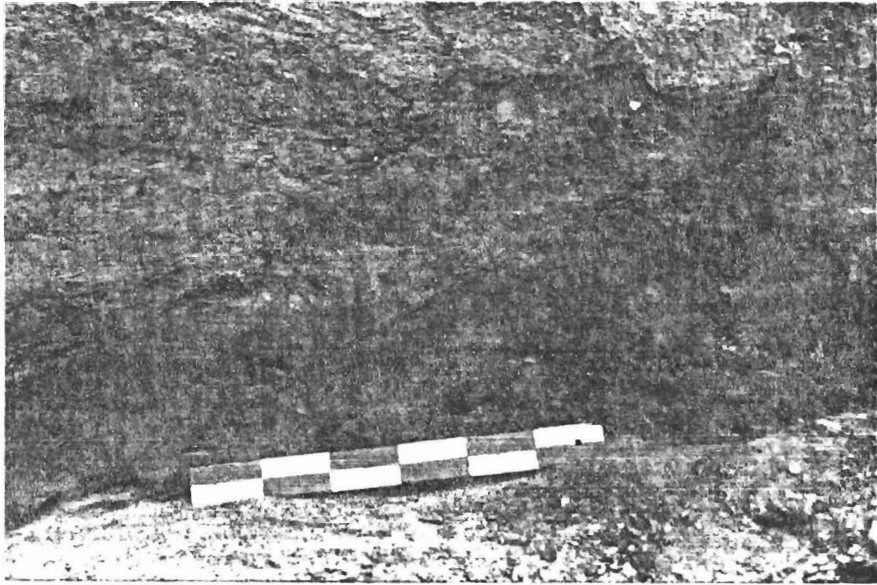
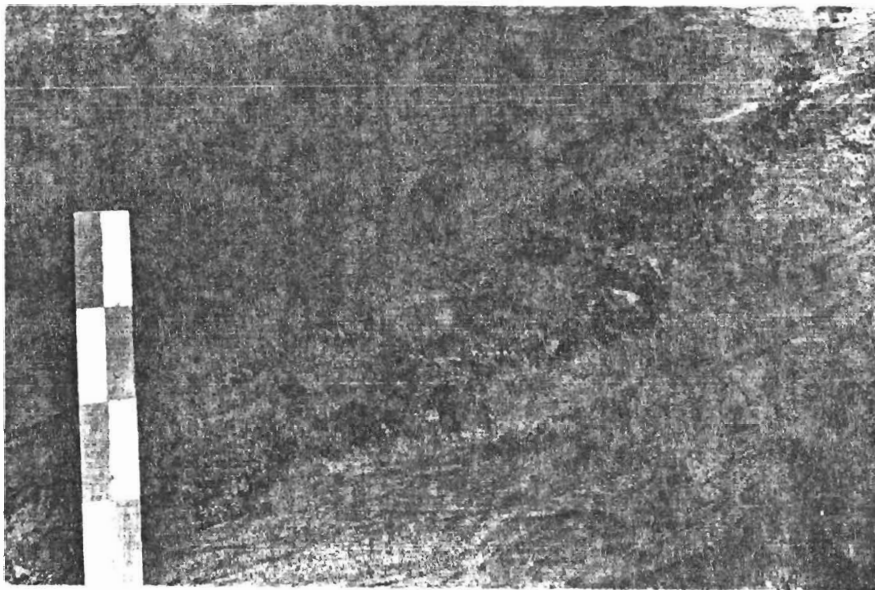


Figure 2.24. Brittle fracture in unit C.

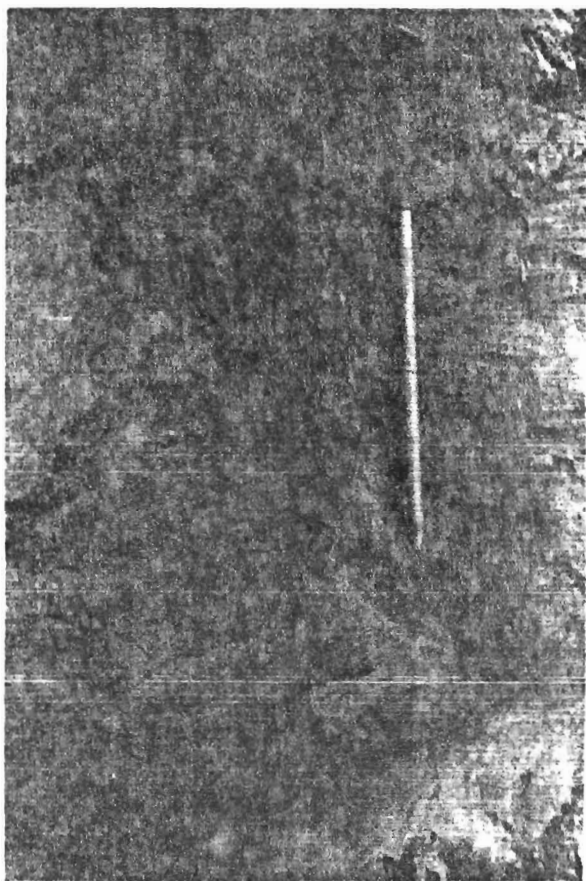
A



B



C



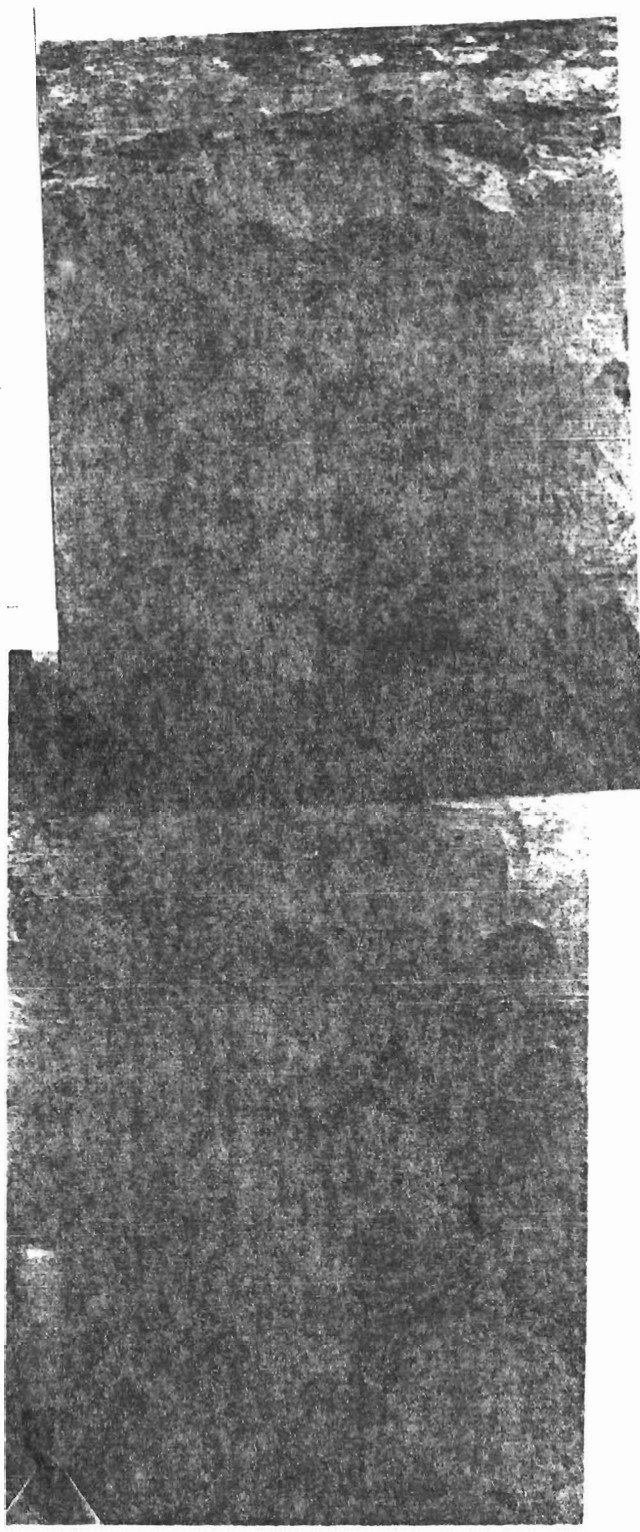
structures are quite common. For example, convolute laminations are always found with other water release features.

This type of lamination can also be found to overlie St and, more often, Sr lithofacies. Figure 2.25 illustrates a case where Sr grades into convolute laminations. Also shown in this figure are the Sr and St lithofacies. Sr is more common than St; St has only been documented in limited places in unit C and always grades upwards into Sr. The latter facies is characterized by both type A and type B climbing ripples. Type A climbing ripples (which may or may not be underlain by St) from the base of a sequence and exhibit a gradation upwards into type B. The transition from A to B is marked by an increasing angle of climb followed by deposition on the stoss side of the ripple form.

Sequences such as that depicted in figure 2.25 are rather uncommon, due to extensive deformation caused by water release. These sequences are normally about 35cm in thickness, although that in figure 2.25 is approximately 2.0m thick, and are truncated by Sd units or may have a capping layer of Dmm. The sequence in figure 2.25 is the only one capped by an Fl lithofacies.

A further ramification of the lack of preserved St and Sr sequences is with regards to palaeocurrent directions. Since the small number of directions that could be obtained from the preserved units would not produce a

Figure 2.25. Gradation from St to Sm. In this photo-mosaic, St occurs at the base and an upwards progression follows with Sr (type B then type A climbing-ripples) grading in Sd then Sm. Note the presence of the dish-and-pillar structure. The sequence is capped by an Fl/Fm layer as shown in the top part of the photograph. Trowel for scale.



significant result, detailed palaeocurrent analysis was not attempted. However, where possible, palaeocurrents were observed to have flowed roughly south to southwest.

Gradational contacts are also present between Sh and St, and Sm and Sr. Figure 2.26 (a) shows the transition of Sh to St. The thickness of Sh is of note in this instance. Sh is usually limited to such occurrences and is thus only of very minor importance in unit C. The transition from Sm to Sr is illustrated in figure 2.27. An important point with respect to figure 2.27 is that, along with apparent coarsening upwards from Sm to Sr, the sequence closely approximates what Reineck and Singh (1980) term alternating bedding. This feature is formed when the depositional environment experiences alternating periods of current activity and quiescence.

Sd was also found to be in contact with Sm. In figure 2.25, Sm overlies Sd truncating the convolute laminations, thus appearing to have been deposited after expulsion of the water from Sd. However, in this instance, grain size appeared to decrease from Sd to Sm; therefore the increase in fines may have decreased the permeability of Sm sufficiently so as to allow vertical movement of water, resulting in horizontal movement. More commonly, the transition between Sd and Sm is gradational: the gradual decrease in the presence of laminae due to the increasing amount of homogenization of the sediment from water release.

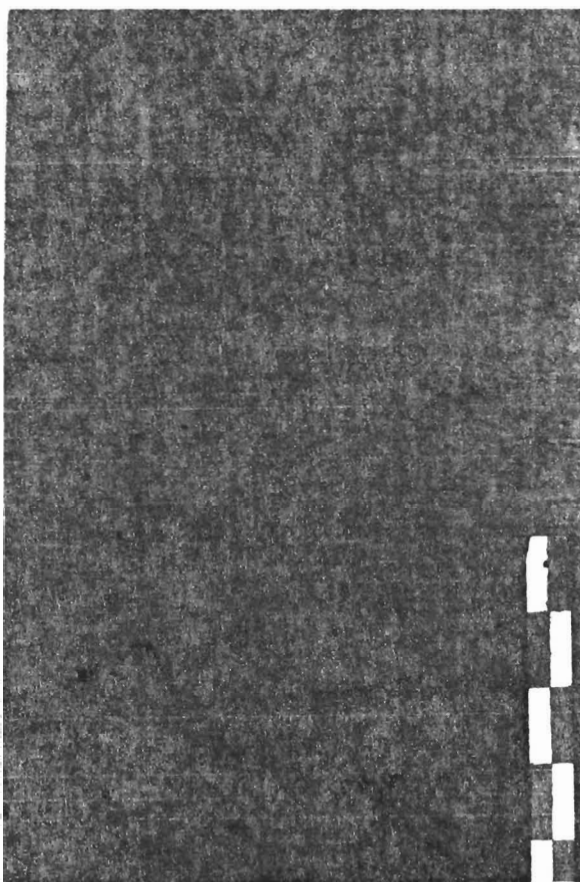
Figure 2.26. Gradational lithofacies for scale.

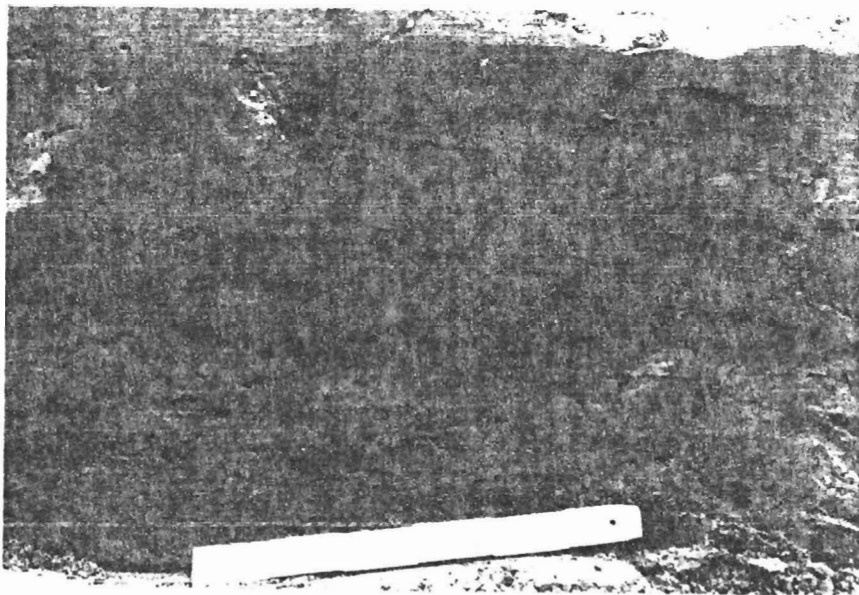
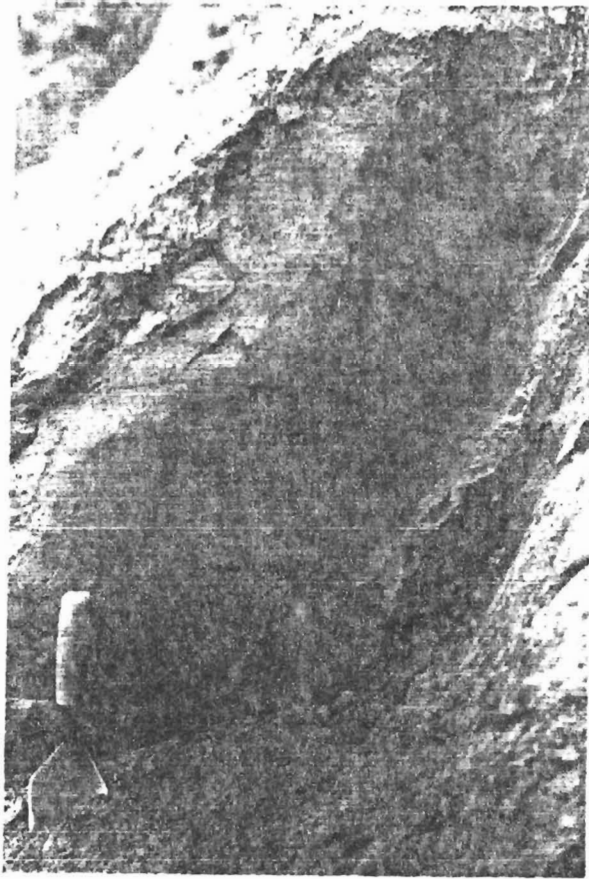
- (A) Finely laminated Sh overlying Dmm layer grading into St.
- (B) Sr and St. Transitions between type A climbing-ripples and small scale trough crossbedded sands.
- (C) St grading into Sr. In this instance, both types A and B climbing-ripples are present.
- (D) Sr lithofacies showing type B climbing-ripples grading into type A. Note increasing angle of climb.
- (E) Close-up of type A climbing-ripples in unit C.

A



B





E



Figure 2.27. Sm to Sr.

(A) Apparent coarsening upwards sequence. Note incipient ripple forms just above clay seams.

(B) Close-up of Sr in (A).

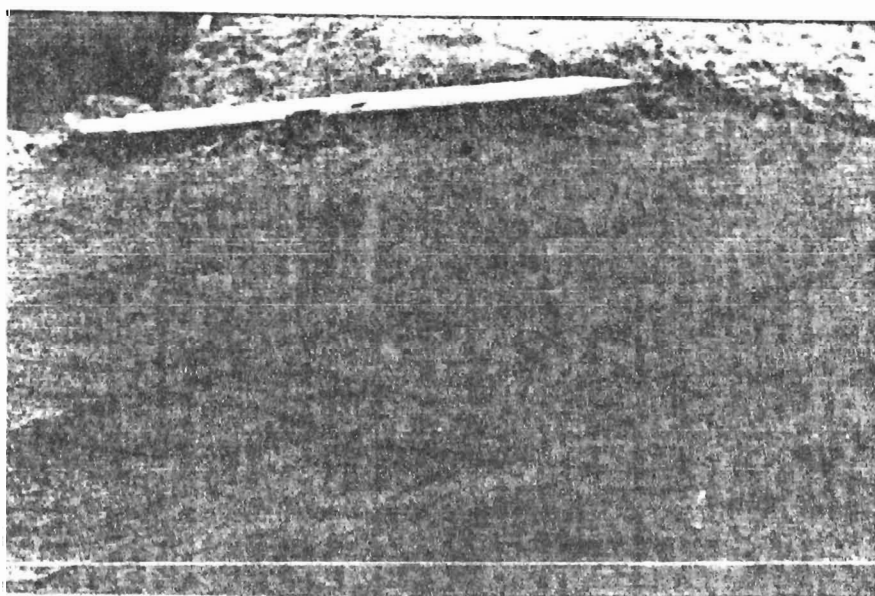
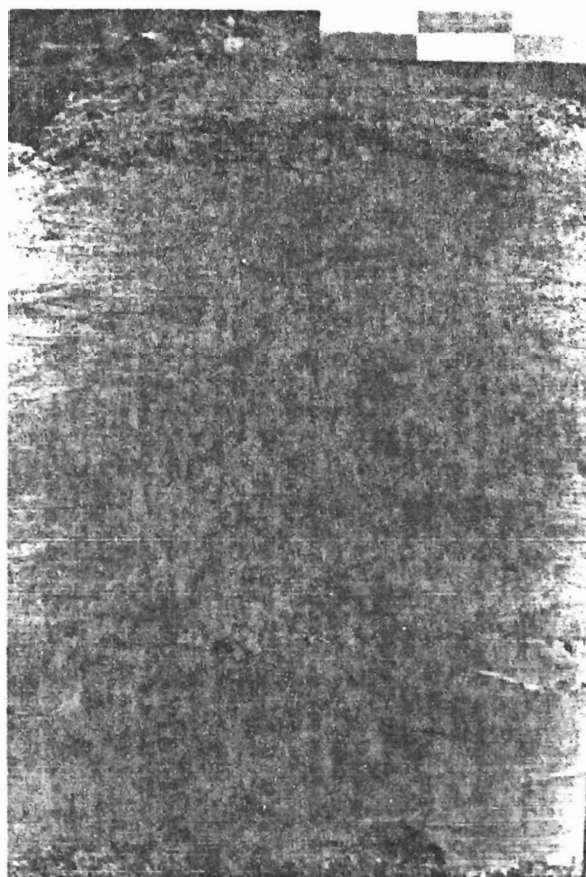


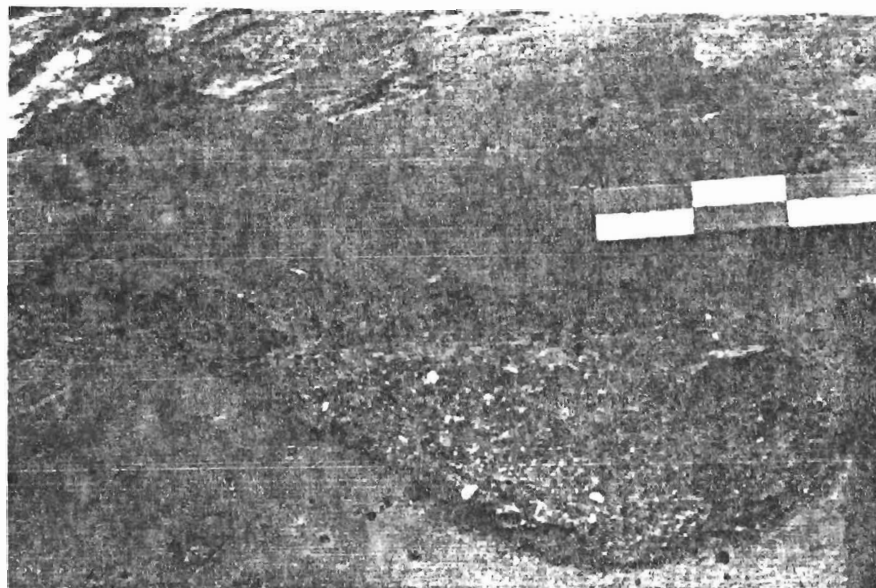
Figure 2.28. Gm lithofacies.

(A) Small-scale matrix-supported channel infill.

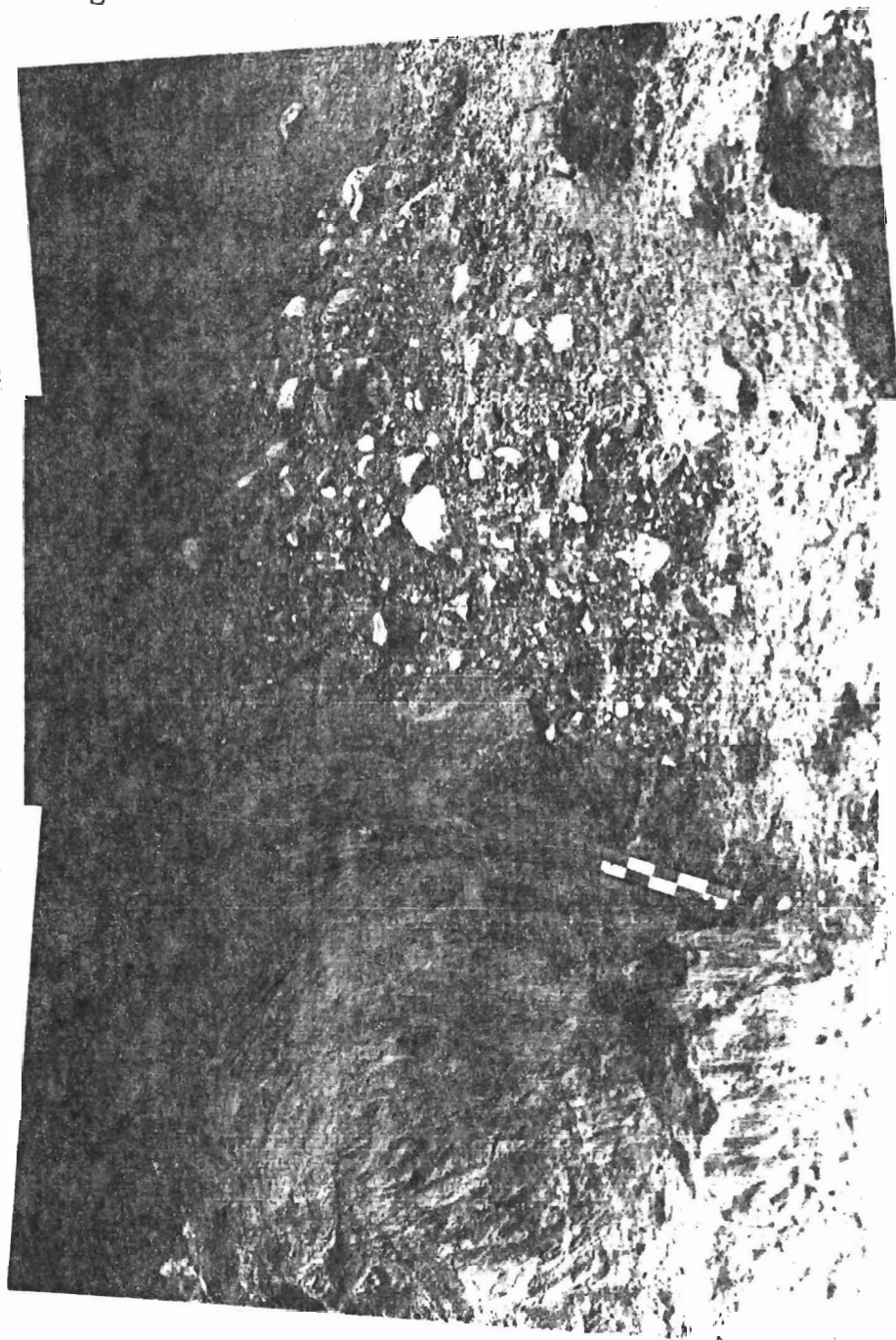
(B) Channel infill deformed by water-escape. Note difference in grain size between (A) and (B).

(C) and (D). Lenticular Gm in association with Sh. Some beds appear to be inversely graded.

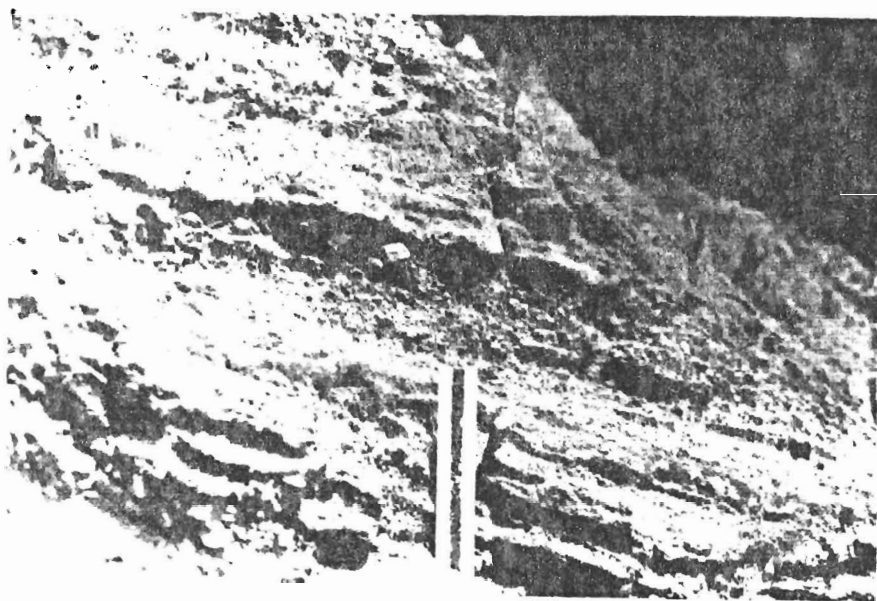
A



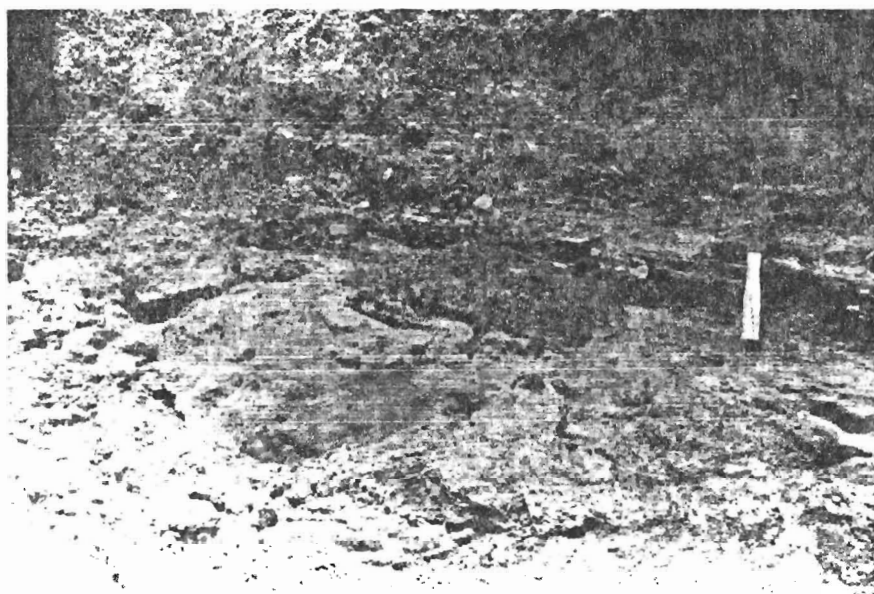
8



C



D



It should be noted that the contact relationships mentioned above refer to facies transitions within beds. Individual beds are about 30cm to 35cm in thickness, with a few exceptions. Contacts between these units are sharp and unconformable. Water escape can either affect the bed internally or a number of beds. From this it can be clearly seen that water release was occurring continuously through time and to varying degrees.

Grain size analysis (fig. 2.4) has indicated that these lithofacies are silty sands. This is, with a few exceptions, consistent throughout irrespective of lithofacies.

As previously mentioned the gravel and diamict lithofacies are of minor extent in unit C. Gravel facies Gm and diamict facies Dmm and rarely Dmm(r)/Fmd are present. Gm is usually in channel-fill sequences or as a lag deposit within sand-filled channels which occur in the upper 2.0 to 3.0m of unit C (figs. 2.23(a), 2.28). Channels are usually quite small: $\leq 1.5\text{m} \times 0.5\text{m}$, with few exceptions. Gm can also form in association with Sh as lenticular layers (figs. 2.23(c) and (d)). Stratification and clast imbrication are absent. Clasts are well rounded, non-striated, and of local origin and $\leq 5.0\text{cm}$. Coarse sand usually forms the matrix support. With the exception of that depicted in figure 2.29(b), soft-sediment deformation does not usually appear to affect Gm.

Contact relationships between Gm and the underlying sand lithofacies are invariably sharp and unconformable. Any laminations are truncated by the gravels. Upper contacts are also sharp. Therefore, the depositional mechanism responsible for Gm must have been sudden, at high velocities, followed by a rapid cessation of activity.

Dmm is not uncommon in unit C. Thicknesses are variable but are normally $\leq 2.0\text{m}$ and can be continuous over distances of up to 10.0m to 15.0m at the maximum (fig. 2.22 (b)).

Unlike that shown in figure 2.22 (b), the most common occurrence of Dmm is at the upper contacts of dish and pillar structures. The reason for this is due to the fine-grained nature of Dmm: because of its grain size, Dmm will have a low permeability and thus will not allow water to pass through. Over time confining pressures will build to a point where water will migrate (horizontally) to a locality of weakness and puncture Dmm, forming the dish and pillar structure (Lowe, 1985).

Contacts between Dmm and the underlying sands are sharp and unconformable. Figure 2.29 illustrates the contacts. As is readily seen, truncation of laminae is evident. Upper contacts appeared to be unconformable initially. However, figures 2.26 (a) and 2.29 (b) and (c) show development of Sh lithofacies ($\sim 0.5\text{cm}$ thick) overlying Dmm. This appears to be quite fine-grained

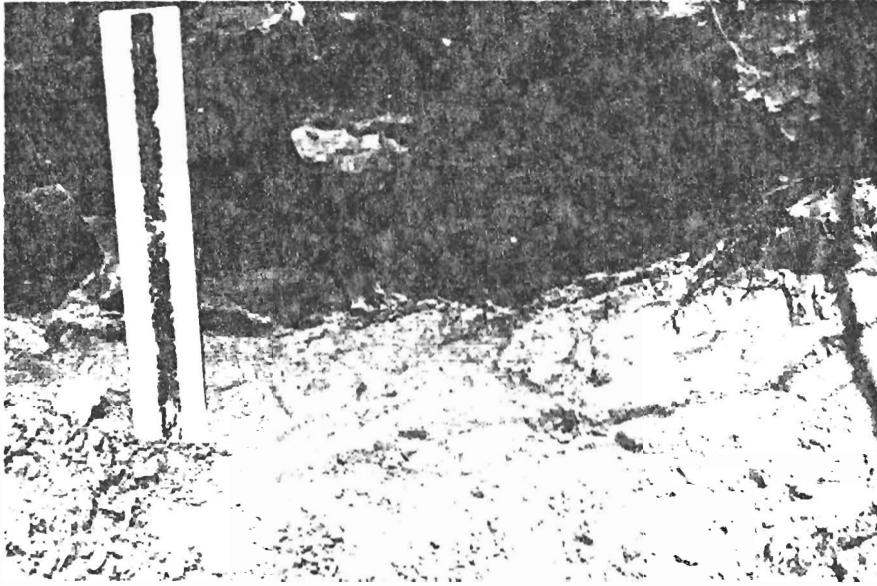
Figure 2.29. Dmm lithofacies.

(A) Sharp basal contact between Dmm and Sm.

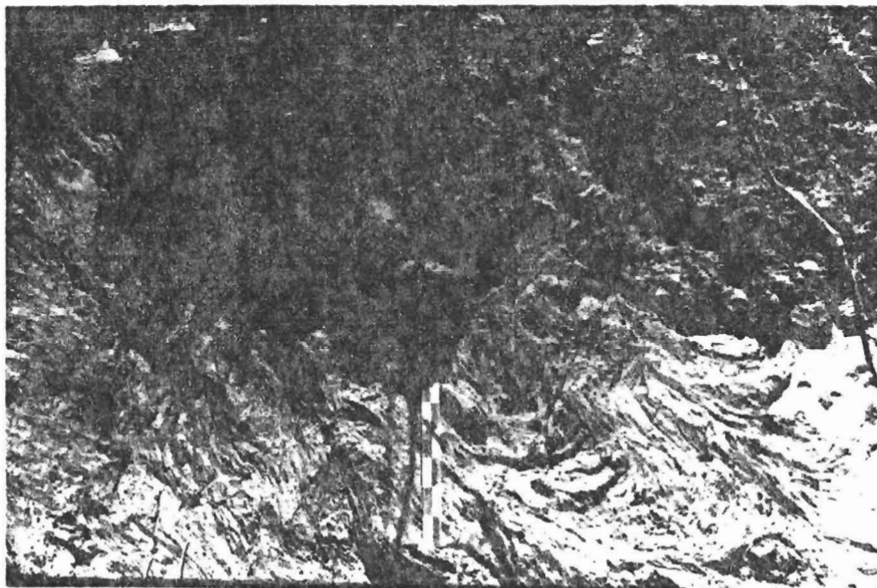
(B) Truncated Sd laminations by Dmm. In both (A) and (B) it can be readily seen that these contacts are quite different to those sand/diamict contacts of unit A.

(C) Close-up of a thin Dmm layer. Notice how the upper part of the Dmm layer had been deformed into a ripple shape. Furthermore, convoluted sands occur within this "ripple" form.

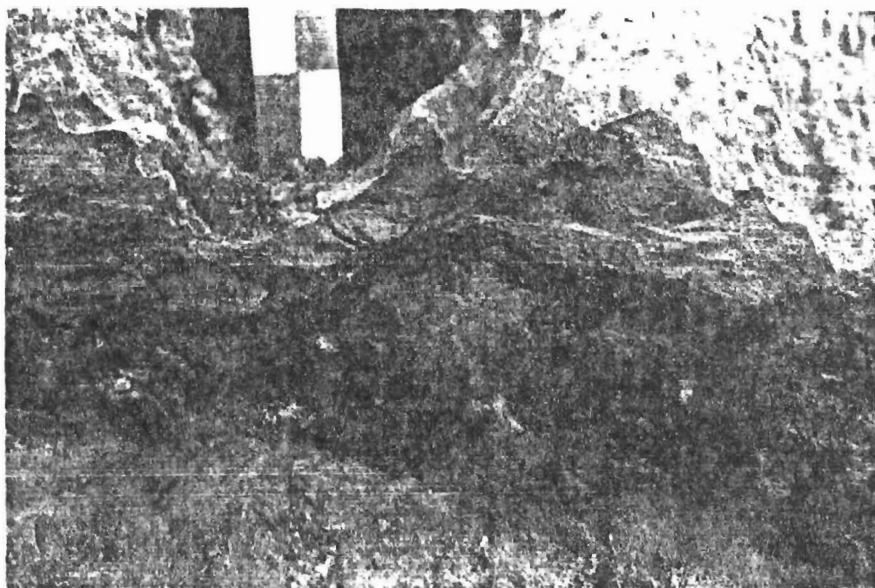
A



B



C



and coarsens upwards into St or Sr. Although the process responsible for depositing Dmm probably waned rapidly, remnants of this activity may have been responsible for the deposition of Sh.

From observation of figure 2.29, it can be seen that Dmm is quite similar to its occurrence in unit B. Indeed, physical properties have not changed between units B and C. Thus, it is reasonable to assume that the mechanism of deposition has not changed either.

An interesting phenomena is pictured in figure 2.29 (c). Along the upper contact with St, ripple-shaped forms are developed. In between the ripple-form and the Dmm layer, Sd is present and appears to reflect the presence of turbulence. Deformation of rippled and crossbedded sands has been documented by Allen and Banks (1972). They attribute folded crossbeds to the drag force imparted to the substrate by an overflowing current. This current is not of sufficient shear velocity to erode its bed; however, enough shear is present to result in deformation of the bedforms.

A similar process is envisaged for the features depicted in figures 2.29 (c). Since Dmm will be near saturation point following deposition, erosion of this sediment should be fairly easy by an overflowing current. However, this was probably of low velocity as indicated by Sh. The presence of Sh may have been

enough to initiate compaction of Dmm, due to overburden pressure and subsequent water release. Upon compaction and de-watering, along with increasing overflowing current velocities as indicated by Sr, entrainment of Dmm ensued but since shear velocity was not enough to fully erode Dmm, deformation into a ripple-shape resulted.

Like unit B, unit C has suffered the effects of pedogenesis. The upper 1.5m is heavily stained in places and the destruction of certain physical properties is suspected. Calcareous concretions have formed at various depths, within the upper 2.0m, sporadically throughout the unit.

Chapter 3

Depositional Mechanisms

3.1 Unit A

In determining the depositional mechanism responsible for unit A sedimentation, several considerations must be taken into account. Unit A is composed of, by and large, a uniform sediment (Dmm) that lacks vertical and lateral changes and is devoid of continuous interbeds. Deformation style is of high shear and low strain rate, capable of macro-scale deformation. Finally, the presence of the sand inclusions, which were formed by entirely different depositional processes than Dmm.

The uniformity of unit A indicates that the depositional environment remained constant throughout the time of deposition and that deposition and deformation were closely related. The sand blocks, however, do represent a different depositional mode than Dmm; due to their discontinuity, their deposition marks only an intermittent and very localized change in depositional style. That Dmm remains unchanged above and below the sand inclusions attests not only to the continuity of deposition, but that the sand lithofacies were probably formed synchronously with Dmm and in the same depositional environment.

These considerations, and the observational data expressed in chapter 2, are fundamental in the determination of process(es) and environment(s) of deposition. These processes are highly variable within the glacial realm (see Sugden and John, 1976, for example) and a detailed discussion of these environments would be far beyond the scope of this thesis. For the sake of brevity, only two environments will be discussed with regards to unit A sedimentation: subglacial and proglacial. Glacio-lacustine or glaciomarine sediments will not be entertained here since their characteristics are quite different than those exhibited by unit A. This will become apparent later in the chapter.

The sedimentological characteristics of the tills and diamicts found subglacially and proglacially are given in table 3.1. These are grouped into lodgement and meltout tills and gravity flows, the former two reflecting the subglacial environment, the latter, proglacial. Two cautionary notes: certain authors feel that subglacial sediments should be subdivided further than what is shown in table 3.1. Examples of such segregation are deformation and lee-side tills. These are supposed to represent sediment comprised mostly of shattered bedrock and till deposited in the lee of a bedrock obstruction, respectively. In the author's opinion, these "tills" are the result of lodging or meltout and should thus be classified as such. Secondly, proglacial

Table 3.1. Characteristics of Lodgement, Meltout Till
and Gravity Flows.

Lodgement	
A) Derivation and Deposition	<ol style="list-style-type: none"> 1) Subglacial derivation. Reflection of local bedrock sources. Transported in traction or suspension layers of basal or lowest englacial zones. Debris becomes more "exotic" in content with increasing distance from basal layer. 2) Deposition due to frictional retardation against bed.
B) Sub-till features	<ol style="list-style-type: none"> 1) Striated bedrock parallel to ice flow 2) Shattering and rafting of rockhead. Transport of rafts and deposition within till. Incompetent lithologies will become folded and attenuated. 3) Intrusion of till into bedrock joints 4) Penetrative deformation, remobilization of sediment beneath till due to overriding of ice. Folding, faulting and thrusting of permafrozen sediments into "stacks". 5) Shearing at contacts of individual units
C) Till features	<ol style="list-style-type: none"> 1) Consists of gravel, sand, silt/clay with varying amounts of each 2) Clasts dispersed through a finer, supporting matrix 3) Clasts show a preferred elongation parallel to ice flow. Clustering may occur. 4) "Flat-iron" shaping of clasts. May form in boulder pavements 5) Massive at macro levels. At the micro-scale, structures indicative of shear present: foliation, shear planes, lineations at bed contacts, "smudges" of incompetent lithologies due to mechanical breakdown from shear. 6) "Cut and fill" channels deposited as sands and gravels from subglacial melt-water streams. Subglacial pondings will deposit laminated silt/clays. 7) Till extending into channel due to bank failures. 8) Upper contacts of channels exhibit erosion and deformation from overriding ice. 9) Subvertical joint orientated either parallel to ice flow or reflecting those in bedrock.

- D) Landforms 1) Drumlinized or fluted terrain, eskers present. "Lee-side tills" on the lee of bedrock highs, rock cored drumlins *

Meltout

- A) Derivation and Deposition
- 1) Subglacially derived, similar to lodgement. Transported in subglacial or englacial zones.
 - 2) Debris is slowly melted out of ice mass under confined conditions such that modification of primary structures does not occur. May be melted out of subglacial or englacial position

- B) Sub-till features
- 1) Similar to lodgement till since the ice depositing this material was initially active and has stagnated. Because of this situation meltout should theoretically be underlain by lodgement till.
 - 2) If underlain by unconsolidated sediment, the contact will be transitional, especially if lodgement substrate.

- C) Till features
- 1) Consists of gravel, sand, silt/clay to varying degrees. Maybe better sorted than lodgement.
 - 2) Clasts dispersed through finer, supporting matrix.
 - 3) Clasts show a preferred orientation parallel to ice flow
 - 4) Upper contact may be marked by faceted boulders forming a boulder pavement
 - 5) Massive to layered at macro-levels. Dependent upon degree of preservation of englacial banding
 - 6) Folding, stacking of thrust slices common.
 - 7) Shear planes, foliation development and joint development also common
 - 8) Incorporation of unlithified sediment blocks
 - 9) Upper contacts with lodgement till (derived from over-riding active ice) sharp, unconformable, exhibit shear phenomena

- D) Landforms 1) Forms as both longitudinal and transverse elements. May also form in "till plains"

Flow Till (Gravity flows)*

- | | |
|------------------------------------|---|
| A) Derivation
and
Deposition | 1) Supra-, en, sub-, or proglacial derivation. May be the result of an agglomeration of all those sources.
2) Deposition from down slope movement due to gravity |
| <hr/> | |
| B) Sub-till
features | 1) If sediment underlying is soft, material may deform by shear. Development of water-escape structures. Rafting may occur. If consolidated, modification of underlying sediment will be negligible. Either case, sharp, non-erosional contacts |
| <hr/> | |
| C) Till
features | 1) Consists of gravel, sand, silt/clay with varying amounts of each.
2) Clasts may be dispersed through a supporting matrix. Clast supported diamicts may also be present
3) Clasts show a poor fabric; dispersed fabrics are common.
4) Massive, graded, and stratified units occur. Grading can be inverse (distribution or coarse-tail) or normal. Flow structures are very common
5) Lower part of unit may exhibit shear if a non-deforming plug was transported within flow. If whole unit was in shear then shear zone will not form.
6) Presence of interbedded units of fluvio-glacial or glaciolacustrine origin.
7) Deposit will resemble an interbedded sequence of flows and associated fluvio-glacial and glaciolacustrine units. |
| <hr/> | |
| D) Landforms | 1) Hummocky topography will result with associated eskers and morainic ridges caused by a retreating ice mass |
| <hr/> | |

*The term "flow till" has been recently disputed by Lawson (1981) and Rappol (1983) in that since these diamicts have been re-deposited and hence modified from their original state, the term till is inappropriate. In agreement with Lawson (1981) and Rappol (1983), the author will continue to term these as gravity flow or mass flow deposits.

after

Boulton	(1970a,b)
Hampton	(1972)
Banham	(1975)
Shaw	(1977, 1979)
Berthelsen	(1978, 1979)
Kruger	(1979)
Lawson	(1979, 1981)
Menzies	(1979)
Eyles <u>et.al.</u>	(1982)
Rappol	(1983)
Thomas	(1984)
Eyles and Miall	(1984)
Menzies and	
Eyles	unpublished

diamict are referred to as gravity flows. Recently, Lawson (1981, 1982) and Rappol (1983), have stated that since diamicts deposited proglacially are by gravity flowage as opposed to direct involvement by the ice, these sediments are not true tills. Therefore, in keeping with this, the author will only refer to sediments deposited directly by ice as tills.

As can be seen in table 3.1, gravity flows occur in association with sediments from fluvial and lacustrine environments, common in the proglacial zone of a stagnant or retreating ice mass. Lawson (1979; 1981) has documented the sedimentation patterns in the marginal zone of the Matanuska Glacier in southern Alaska, and states that resedimentation processes by gravity flow are the most dominant form of deposition. Debris may undergo reworking several times in one season. Flows have been characterized on the basis of several physical parameters, which vary with water content, into four types (Lawson, 1981; 1982). These characteristics are presented in table 3.2 and illustrated in figure 3.1. These flows have similarities with some of the more viscous types of flows common in delta fans and continental shelves, as discussed by Lowe (1979) and Nardin et.al. (1979). Examination of tables 3.1 and 3.2 and figure 3.1 readily demonstrate that the properties of gravity flows with their associated fluvioglacial and glaciolacustrine deposits and landforms are quite different than the

Table 3.2
Sediment flow properties (after Lawson, 1981, 1982).

Deposit	Texture type 1) mean (ϕ) 2) ϕ (ϕ)	Internal organization			Surface forms	Contacts and basal surface features	Penecontem- poraneous deformation	Geometry and maximum dimensions
		General	Structure	Pebble fabric				
Melt-out till	Gravel 1) to 6 sand. 2) 1.8 to 3.5 silt; silty sand; sandy silt.	Clasts randomly dispersed in fine-grained matrix.	Massive; may preserve individ- ual or sets of ice debris strata as subparallel laminae and lenses.	Strong; unimodal parallel to local ice flow; low angle of dip.	Similar to ice surface; may be de- formed or faulted.	Upper sharp, may be trans- itional; sub- ice probably sharp.	possible inter- nally and in overlying mat- erial during ice melt.	Sheet to dis- continuous sheet; several km in area, 1-6m thick.
Sediment flow deposits	Gravel-sand- silt, sandy silt, silty sand 1) 2 to 3 2) 3 to 4	Clasts disp- ersed in fine- grained matrix.	Massive	Absent to very weak; vertical clasts.	Generally planar; also arcuate rid- ges, second- ary rills and dessic- ation cracks	Nonerosional -conformable contacts; contacts sharp; load structures.	Possible sub- flow and mar- ginal deform- ation during and after deposition.	Lobe: 50x20m, 2.5m thick
	Gravel-sand- silt, sandy silt. 1) 2 to 3 2) 3 to 4	In plug zone, clasts disp- ersed in fine- grained matrix.	Massive; intra- formational blocks common.	Absent to very weak; vertical clasts.	Aucuate ridges; flow lineations, marginal folds, mud volcanoes, braided and distributary rills.	Nonerosional conformable contacts; contacts in- distinct to sharp; load structures.	Possible sub- flow and mar- ginal deform- ation during and after deposition.	Lobe: 30x20m 1.5m thick; sheet of coalesced deposits.
		In shear zone, gravel zone at base, upper part may show decrea- sed silt, clay and gravel con- tent; overall, clasts in fine- grained matrix	Massive; deposit may appear lay- ered where shear and plug zones are distinct in texture.	Absent to weak; bimodal or multimodal; vertical clasts.				
	Gravelly sand to sandy silt 1) -2.5 to 2.5 2) 3.5 to 2	Matrix to clast dominated: lack of fine-grained matrix possible; basal gravels.	Massive; intra- formational blocks occas- ionally.	Moderate, multi- modal to bimodal, parallel and transverse to flow direction.	Irregular to planar; sing- ular rill development, mud volcanoes	Nonerosional conformable contacts; contacts in- distinct to sharp.	Generally absent; possible sub- flow deform- ation on liquefied sediment.	Thin lobe or fan wedge: 30x65m, 3.5m thick, less often sheet of coalesced deposits.
	Sand, silty sand, sandy silt 1) 3.5 2) 2.5	Matrix dominated except at base where granules possible.	Massive to graded (dist- ribution, coarse-tail)	Absent	Smooth, planar: mud volcanoes possible.	Contacts conformable; indistinct.	Absent	Thin sheet: 20x30m. 0.3 thick. Fills surface lows of irregular size and shape.

increasing water content

Attribute	Flow Type			
Morphology	Lobate with marginal ridges; non-channelized.	Lobate to channelized	Channelized	Channelized
Channel-wise Profile	Body constant in thickness with planar surface: head stands above body; tail thins abruptly upslope	Body constant in thickness with ridged or planar surface: head stands above body; tail thins abruptly upslope.	Flow thins from head to tail; irregular surface.	Flow constant in thickness; planar surface
Thickness (m)	.01 to 2.0	.01 to 1.6	.03 to .6	.02 to .15
Bulk Water Content (dry wt %)	11 to 17	17 to 24	25 to 33	35 to 70
Bulk Wet Density (gm cm ³)	2.0 to 2.6	1.9 to 2.15	1.8 to 1.95	1.1 to 1.7
Bulk Mean Grain size (mm)	2.0 to .3	1.9 to .1	.5 to .06	.1
Surface Flow Rates (cm/sec)	.1 to .5	.2 to 5	15 to 125	1 to 200
Typical Length of Flow (m)	10 to 300	10 to 100	50 to 400	30 to 400
Surface Shear Strength (kg/cm ²)	4 to 1.5	.6 or less	Not measurable	Not measurable
Grain Support and Transport	Gross strength; shear in thin basal zone	Gross strength in rafted plug, traction, localized liquefaction and fluidization grain interactions and reduced matrix strength in basal and marginal shear zones	Traction, grain dispersive pressures, possibly liquefaction fluidization. transient turbidity. shear throughout	Liquefaction, some traction, grain interactions, differential shear

Central part of flow

Maximum length may reflect topographic conditions of terminus region

Figure 3.1. Illustration of flows as classified in table 3.2.

- A) crosssections of flows, parallel and transverse to movement. Note changes in physical parameters.
- B) idealized internal structures of flows with varying water contents.

Unit 1: massive to normal grading, texturally heterogeneous, gravel within traction zone.

Unit 2: massive, texturally heterogeneous, large grains absent.

Unit 3: massive, but has zones of texturally distinct or structured sediment, vertically oriented clasts common.

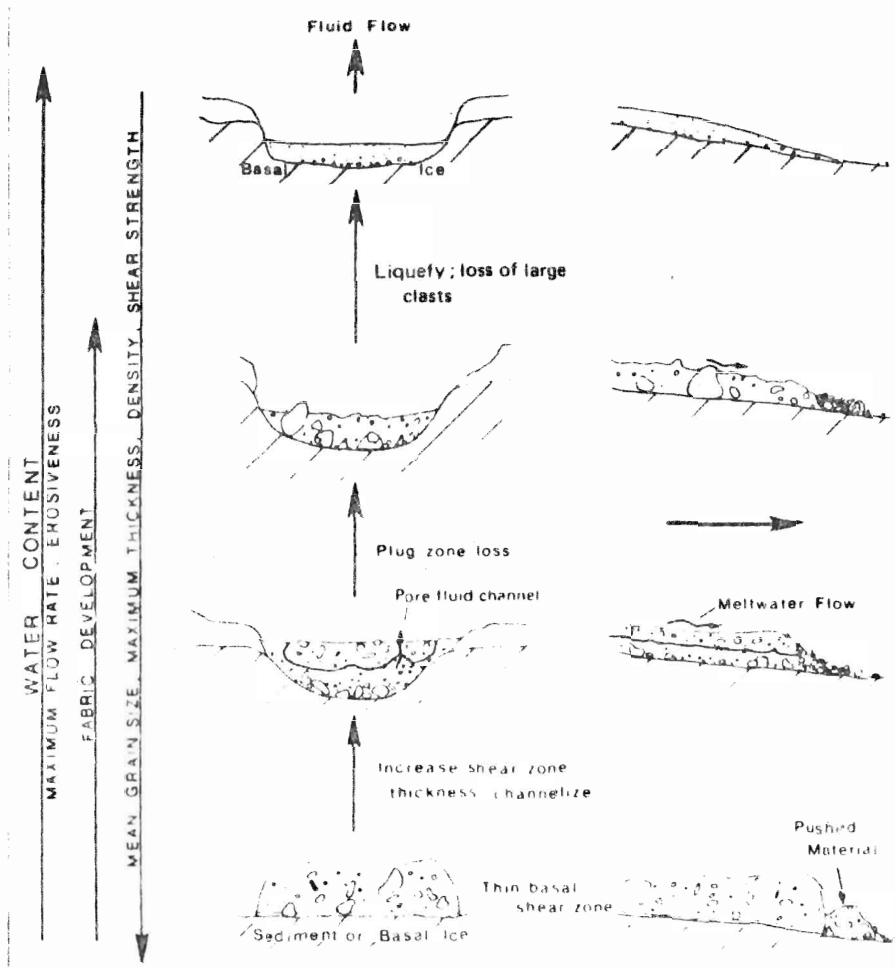
Unit 4: massive, fine-grained matrix similar to unit 2, lacks coarse clasts.

Unit 5: horizontally laminated sands and silts deposited by meltwater.

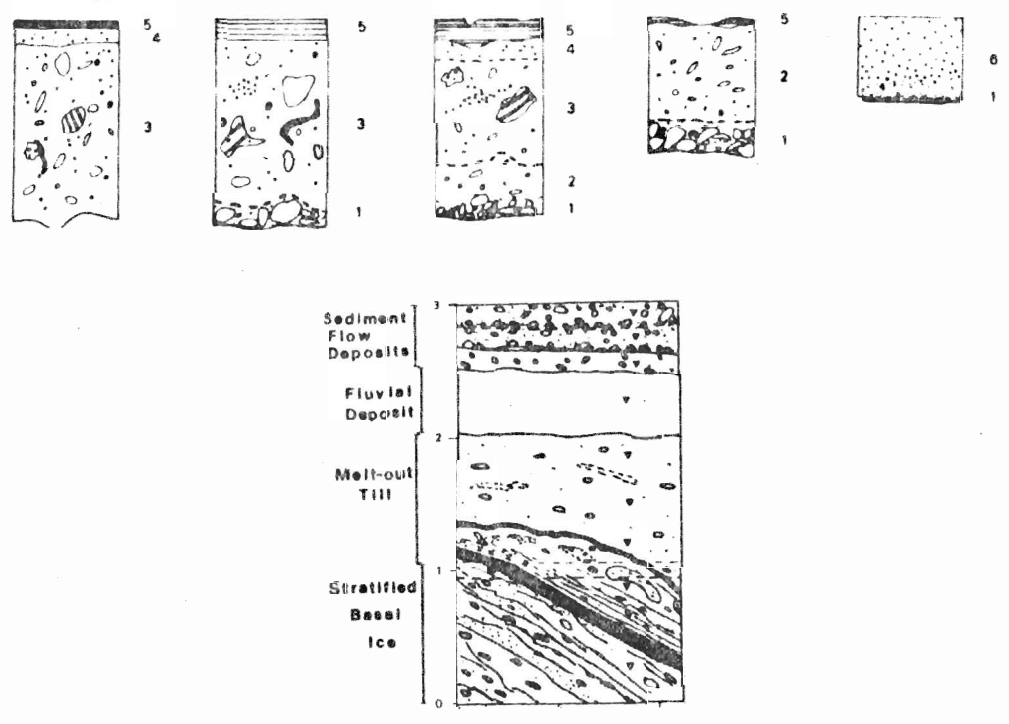
Unit 6: massive or normal to coarse-tail grading, silty sand.

- C) typical proglacial sequence at the Matanuska Glacier.

(Lawson, 1981)



INCREASING WATER CONTENT OF SOURCE FLOW



characteristics exhibited by unit A. Therefore deposition by sediment gravity flow in a proglacial environment may be discounted.

The features of lodgement and meltout tills that are deposited subglacially, are described in table 3.1. When compared to the characteristics of gravity flows, subglacial tills exhibit far more deformation structures than mass flows both internally and externally ie with respect to intercalated sediments and their substrates. The reason is due to depositional process. Lodgement tills are deposited when the drag force on the debris exceeds the shear stress applied by the ice, whereas meltout tills form due to in situ melting of stagnant, debris-laden ice. If debris-laden ice overrides soft, incompetent bedrock or unlithified sediment, one can see that because of the low strain rates involved, the substrates will become deformed. Examples of bedrock modification by ice are documented by Banham (1975) and Berthelsen (1978; 1979) who show well developed folds and attenuation features exhibited by bedrock floes within lodgement tills in the United Kingdom and Scandanavia, respectively. Deformation of unlithified, albeit permafrozen, sediments is a very common occurrence. Recent cases of this phenomenon are examined by Bluemlle and Clayton (1984), who discuss ice-thrust features formed during the late Wisconsin in North Dakota. Similar work has been performed by Thomas (1984) and Thomas and Summers (1984) in the U.K., and Eire, respectively, who attribute

large scale thrusting and folding of unlithified sediments to be the overriding of proglacial and glacial sediments by an advancing ice sheet. It should be noted that Eyles and Eyles (1984) have developed a model for the Isle of Man sediments discussed by Thomas (1984), and do not invoke the involvement of an advancing ice sheet over a permafrozen sandur but to glaciomarine processes.

The permafrozen nature of the underlying sediments is fundamental to the development of glacio-dynamic structures. Entrainment of the substrate is accomplished by a "freezing-on" process (Boulton, 1970 b): interstitial water within saturated debris is frozen on to the base of the ice sheet and in the process the debris is incorporated into the ice sheet (fig. 3.2). The freezing-on process is governed by the penetration of a freezing front through the debris which is in turn dependant upon the permeability and the porosity of the sediment (Menzies, 1981 a and b). Once incorporated into the basal zone of the ice, the debris will experience intensive shearing, and abrasion from interparticle contacts. From this it is easily seen how incompetent bedrock and permafrozen slices are deformed while in transport.

Deformation need not entail significant transport. Thomas' (1984) model for the development of the Bride Moraine of the Isle of Man serves as an excellent example. The sediments that comprise the Bride Moraine are a series

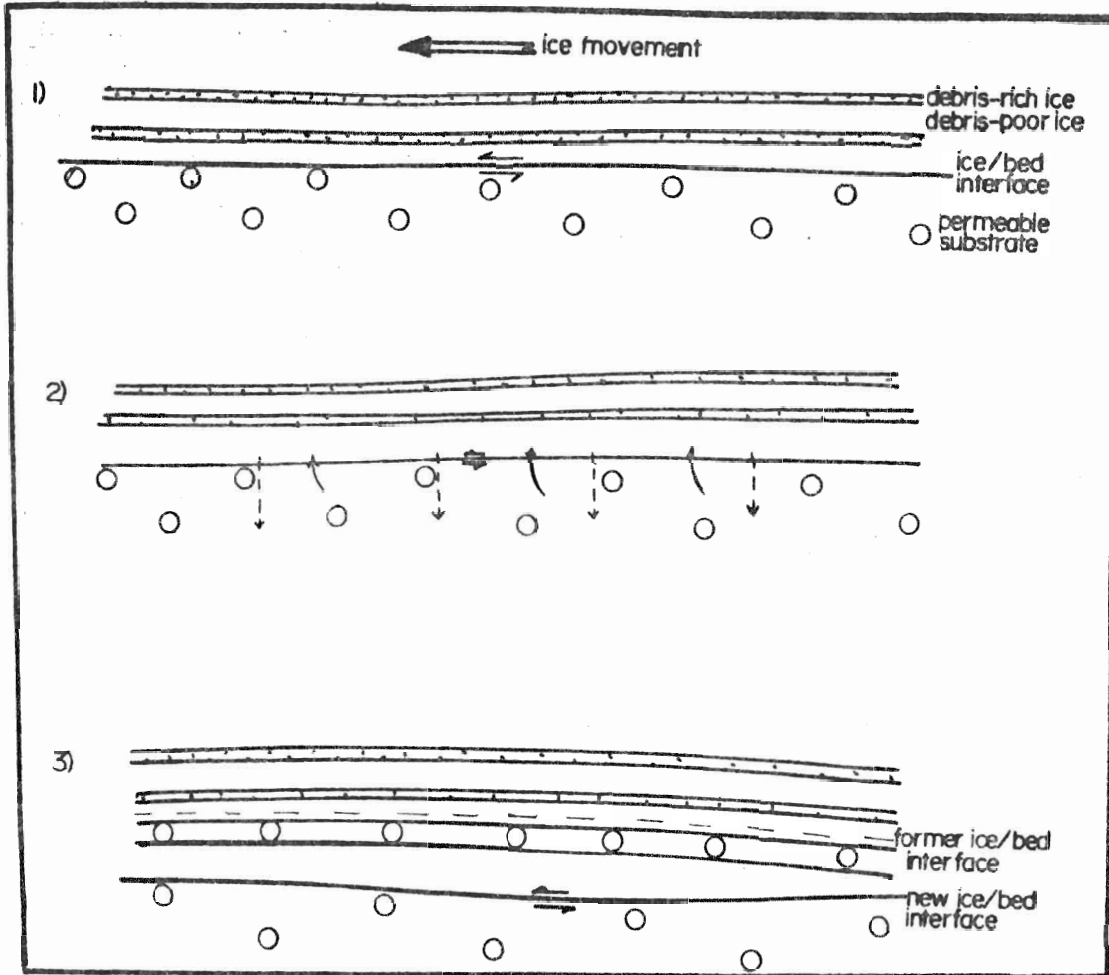


Figure 3.2. Freezing-on process.

- 1) ice moving over a permeable substrate
- 2) interstitial waters freeze to base of ice sheet (indicated by solid arrows) as freezing fronts penetrate basal material (dashed arrows)
- 3) entrainment of debris (after Boulton, 1970b).

of proglacial and glacial sediments which have been deformed into a series of folds and thrust slices. The contrasting lithologies of the sequence caused the differential development of permafrost: sands and gravels were frozen due to their high porosity and permeability whereas the silt/clay-rich sediments were not frozen, not only due to a lower transmissibility but due to the presence of supercooled interstitial water. This resulted in the fine-grained sediments deforming in a more ductile fashion than the coarse-grained sediments, thus allowing the latter sediments to move as a result of the shearing of the fine-grained material.

If silt/clay-rich sediments are not present to facilitate movement, permafrozen materials will internally deform. Permafrozen soils are either ice-rich or ice-poor. In both cases ice forms as a cement; in ice-rich soils ice lenses and veins also occur. In the latter material, ice content is such that any intergranular contact is removed and thus deformation will approximate that of pure ice (Nixon and McRoberts, 1976). Grain to grain contact will result in deformation characteristics somewhat different from that of ice-rich soils. Work by Roggensack and Morgenstern (1978) and Morgenstern et.al. (1980) has shown that in ice-poor silty clays shear stresses necessary for deformation were ≥ 200 kpa, much greater than that for ice-rich soils. Internal angles of friction were found to be similar

to that of the same sediment in the thawed state. This may be due to the growth of ice lenses within the material during deformation. Examination of these features showed displacements and fabric developments.

Although permafrozen sediments deform, the effect of ice cementation is to cause the sediment to behave as a unit. Without the cement of ice, debris particles would be entrained individually, and would suffer abrasion within the basal zone.

Since the substrate of an ice sheet displays deformation due to shear, it is not surprising that lodgement and meltout tills exhibit structures produced by shear. The most documentation on this subject involves that of till fabric. Fabric is basically a lineation developed by elongate grains parallel to the direction of shear (Hobbs et.al. 1977).

The applicability of fabric measurements has been discussed by Young (1969) and Mark (1974). Young (1969) found that over short distances ($\approx 1.0\text{m}$) both lateral and vertical variations in orientations were considerable; furthermore that plunge was highly variable and that no ice movement directions should be inferred from this data. The observations by Young (1969) were made on coarse-grained till (Menzies, personal communication, 1983), thus interactions between grains, although tending to give preferred orientations, may have also hindered in

fabric development. Conversely, Mark (1974) claims that plunge, if statistically significant can be used as a directional indicator. Mark (1974) does state that if plunge is not statistically significant, directions of ice movement should not be attempted.

More recently, however, work on lodgement and melt-out tills by Kruger (1979) and Shaw (1982), respectively, show that fabric measurements can provide an indication of ice movement. Therefore, bearing in mind the limitations expressed by Young (1969) and Mark (1974), fabric measurements can be used, but should be used in conjunction with other directional indicators such as drumlin directions and striations.

A further effect of shear imparted on subglacial debris is the development of lineations and foliations. The importance of these features have recently been pointed out by various workers (e.g. Ehlers and Stephen, 1979; Eyles et.al., 1982; Shaw, 1982; Thomas, 1984). Lineations are normally developed along contacts with intercolated or underlying sediments on the underside of the till. These are usually slickensided or "ribbed" surfaces orientated parallel to the shear direction (Ehlers and Stephen, 1979; Rappol, 1983). Foliations involve the alignment of platy minerals into the plane of shear (Hobbs et.al., 1977). In tills, these may be produced by clay mineral alignments on the microscopic

scale, or on a larger scale, the alignment of all particles, which gives the appearance of cleavage.

Thomas (1984) claims that the latter form can be confused with sedimentary bedding; with regards to the Isle of Man sediments which Thomas (1984) states exhibit cleavage, Eyles and Eyles (1984) interpret this as a stratified diamict or Dms lithofacies.

Development of meltwater channel sediments within subglacial tills is not uncommon. Eyles and Sladen (1981), Eyles et.al., (1982), Shaw (1982) and Rappol (1983) all give examples of these features in both lodgement and meltout tills. The subject of glacier hydrology and the flow of water at the base of an ice sheet has been debated for some time (Paterson, 1981 and references therein). Water can flow in three different forms subglacially: in channels incised upwards into the ice (R-channels), channels within the substrate (N-channels), or in sheet flow. Theoretical analysis has shown that both channeled and sheet flow may occur; however the time duration and extent of channelization is still under contention (Walder, 1982; Weertman and Birchfield, 1983). Sheetflow may be more widespread than previously thought: Shaw (1983) attributes the formation of drumlins to the action of subglacial meltwater flow in sheets.

Meltwater deposits can range from laminated clays, due to ponding, to coarse gravels produced by energetic

stream flow (Eyles, et.al., 1982). Deformation of those sediments is quite common (Shaw, 1982; Rappol, 1983) involving destruction of primary structures, channel geometry and production of sheared contacts, for example. The reasons for and mechanisms of deformation have previously been discussed with regards to permafrost development.

Although deformation is quite common, preservation of primary structures may still occur. Harris and Bothomley (1984) document a present-day case where meltwater deposits have been exposed at the margin of Leirbreen in Norway. These sands and gravels have well developed fore-set beds and climbing ripple sequences which are interpreted as deltaic in origin, were formed englacially within a cavity, frozen and transported along shear planes, without any apparent deformation.

From the following discussion it is obvious that the features found within Dmm strongly suggest a subglacial origin. Clast fabrics paralleling lineation orientations, development of shear structures and the deformation exhibited within the sand inclusions run contrary to a proglacial origin.

Whether Dmm is a lodgement or meltout till is very difficult to determine. A glance at table 3.1 shows that these tills are quite similar although they are formed differently. In modern environments meltout tills are presently forming and are easily identified (Boulton, 1971a;

Shaw, 1977a, Lawson, 1979). Holdersen and Shaw (1982) claim that meltout till is distinguished by the presence of interstratified, unlithified sorted sediments, preferred clast orientations parallel to ice flow, and a till configuration closely approximating englacial debris. The former two characteristics, which are also common in lodgement tills (Eyles et.al., 1982), are found within Dmm.

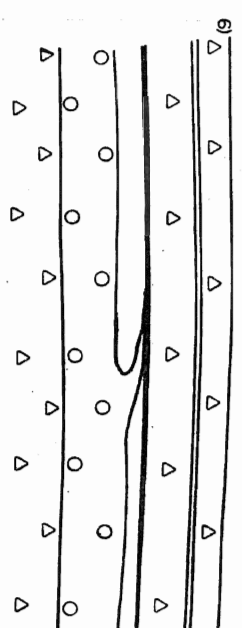
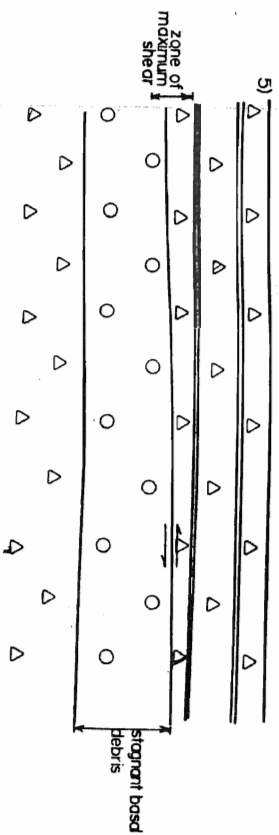
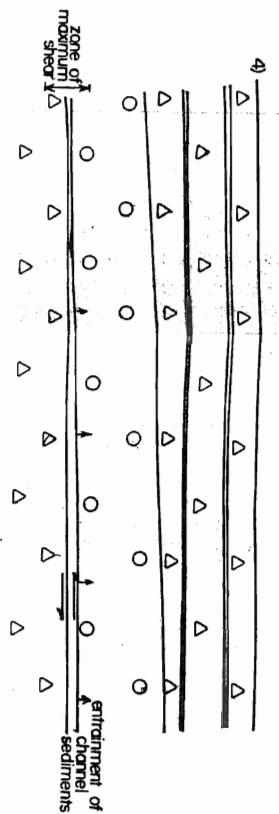
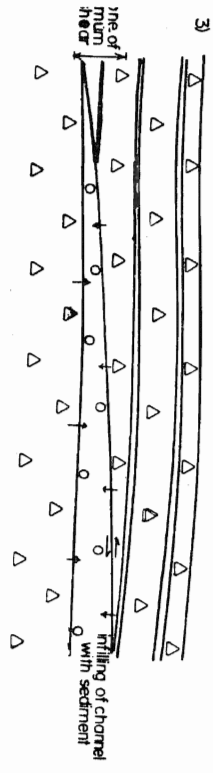
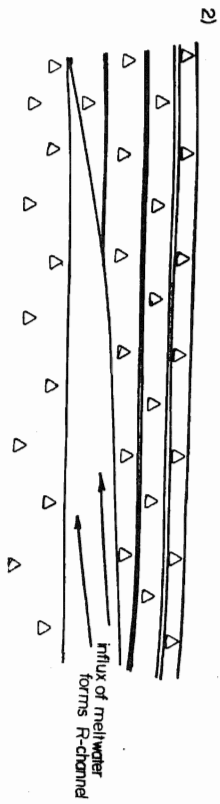
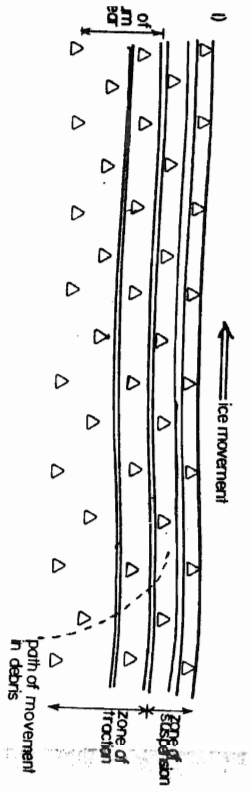
The presence and deformation of the sand lithofacies may hold the solution. Consider an ice sheet advancing over unlithified proglacial sediments. Deformation of these sediments and deposition of lodgement till would ensue. Melting of the ice would produce both englacial and subglacial meltwaters which would tend to channelize along the bed. Depending on discharge, which may have fluctuated to some degree, the water flow may have approximated fluvial conditions or surges of debris-laden water. Channelization would either be in R- or N-channels; R-channels may be more likely since erosion of debris-poor ice would be easier than debris-rich ice. Sedimentation in the conduit would begin to close the cavity. Concurrently, the meltwater sediment would begin to freeze due to the ambient debris temperatures. Closure of the cavity would result in deformation of the upper part of the sand unit, destroying or modifying primary structures. At this point the sand inclusion would be entirely in the frozen state and would undergo freezing to the basal part of the ice

sheet, ie. entrainment from freezing-on processes. Because of the sudden increase in debris contact, the shear strength of the basal zone would increase substantially. This increased shear strength may have exceeded the shear stress applied by the ice, thus resulting in stagnation of the basal layer, and an upward shift in the zone of shearing. The stagnant debris-rich ice would then have experienced deformation from the overriding ice, and would have lodgement till deposited on top of the stagnant ice. This scenario is illustrated in figure 3.3.

Therefore, what is envisaged is a succession of lodgement-meltout processes acting at the glacier bed depositing till and resulting in deformation of the sediments. Initially as the ice advances, lodgement till will be deposited. This will continue until the development of the meltwater channels is initiated. Meltwater channels only mark a localized stoppage in lodgement: where channels do not occur, lodging will continue. Once channels close, lodgement will initiate again, but only briefly. Stagnation of the basal layer and the upwards shift of the zone of shear result in the onset of meltout conditions. In the field, identification of this occurrence, the upper contact of the meltout till/channel sediment association, is very difficult. In places, shear zones are found approximately 10.0 to 15.0m above the sand inclusion till contact, and in others no shear zones exist.

Figure 3.3. Depositional model for unit A.

- 1) Deposition of lodgement till onto a mobile bed.
- 2) influx of subglacial meltwaters results in development of R-channels.
- 3) channels infill with sediment. During this time ambient temperatures will cause sediment to freeze (indicated by arrows showing penetration of freezing fronts).
- 4) freezing-on process results in entrainment of channel sediment.
- 5) stagnation of basal ice due to incorporation of meltwater sediment. Zone of maximum shear shifts upwards.
- 6) deposition of lodgement. Overthrusting of ice will result in deformation of channel sediment.



The shear zones in the former case are probably the melt-out/lodgement contacts, whereas in the latter, the Sd/Dmm contact may be the zone of décollement or the contact may be obscured entirely.

The model explains the deformation of, and structures within, the sand lithofacies. The climbing ripple sequences, common in deltaic and turbiditic environments which experience periodic surges of debris-laden water, reflect the pulsating nature of the waters within the channels. The sheared upper contacts, till injections and convoluted laminations are all products of shear from the overriding ice. The faulting may have two origins. Thrust faults, indicative of compression, were probably formed during deformation by the upper lodgement till. Normal faults, on the other hand, are a type of collapse structure formed when the ice cementing the debris escaped from the till resulting in subsidence.

Large scale structures were the result of entrainment or overthrusting processes. The inclusion in figure 2.8 (a) appears to be deformed to a minor extent on the macro-scale. Closer inspection revealed that the upper part of the inclusion was deformed and separated from the lower undeformed part by a till inclusion. The upper part undoubtedly suffered deformation from entrainment. The lower part escaped untouched due to the pressure of the till inclusion. The till inclusion, being of low permeability, did not

allow penetration of freezing fronts into the lower sands. Shearing, therefore, took place within the inclusion itself, deforming the upper part and just below the till layer as shown in figure 2.11 (d).

The more highly deformed inclusions were probably deformed by the ice overriding stagnant ice. The stagnant ice/sediment mixture would have experienced shear from the over-thrust motions of the upper ice. This resulted in contorting and folding of the channel sediments. Examples of this phenomenon are found in Boulton (1970a) and Shaw (1977a).

Theoretical analysis has suggested that subglacial channels are in existence briefly (Weertman and Birchfield, 1983). The subglacial channels in unit A have climbing ripple sequences preserved, when undeformed. These sedimentary structures have been studied extensively (e.g. Allen, 1971, 1973; Ashley et.al. (1982), and references therein). Flume studies by Ashley et.al. (1982) have shown that a sequence of climbing ripples, ~20cm thick, that closely approximated the natural environment, was deposited in 1.5 to 3.0 hours at flow velocities of 15 to 40cm/s. The sequence depicted in figure 2.9 is approximately 2.5m thick. Applying the findings of Ashley et.al. (1982), this sequence could have been deposited in as little as 20 to 40 hours. How realistic this time duration is, is unknown. However it does indicate that these channels may well indeed be only brief occurrences.

3.2 Unit B

Unlike unit A, unit B is not characterized by the predominance of one lithofacies. Individually, none of the three lithofacies exhibit domination, with one minor exception. Near the ridge, Dmm is present with minor Dmm(r)/Fmd intercalations. Thus, conditions during sedimentation were variable to some extent, and the lithofacies reflect these variations.

The major difference with unit A, along with that stated above, is that unit B does not appear to exhibit any form of glaciotectonic deformation. The lack of sheared contacts, lineations and large scale deformation illustrates that the conditions which prevailed during unit B deposition were totally different from unit A. Therefore, the subglacial environment may be discounted as a possibility for unit B.

Table 3.1 shows that sediments deposited proglacially are highly variable in lithofacies composition. What is not shown in table 3.1 are the other two major proglacial environments: glaciolacustrine and glaciomarine. The glaciomarine environment can be neglected as a possibility since there are no records of these conditions in southern Ontario during the Pleistocene (Prest, 1969). However, since sedimentation processes in large proglacial lakes are virtually the same as those found in the marine

environment (Andrews and Matsch, 1983), reference will be made where applicable.

Terrestrial proglacial sedimentation is characterized by a definitive suite of landforms and sediments, often termed a landform association (Boulton, 1972; Boulton and Paul, 1976). Boulton (1972) and Boulton and Paul (1976) have defined three different landform associations: proglacial, supraglacial and glaciated valley. The proglacial/subglacial system is composed of lodgement till and outwash which take the form of drumlins, fluted and push moraines, kame and kettle and eskers. The supraglacial, which is superimposed on the proglacial/subglacial system, is mostly kame moraines composed of melt-out till and sediment gravity flows. The valley system is made up of medial and lateral moraines and kame terraces of supraglacial till and debris (fig. 3.4).

Although there are some flaws in this approach, since it is based on idealized sequences and does not take local variables into account, it does illustrate which sediments are found in the terrestrial proglacial environment and their associated landforms. Outwash sediments, which Boulton (1972) and Boulton and Paul (1976) state comprise the proglacial zone, are extremely variable. These can be laminated and flow sediments in kettle holes, flows and fluvial deposits within moraines parallel to the ice front and braided stream deposits (Boulton, 1972; Boulton

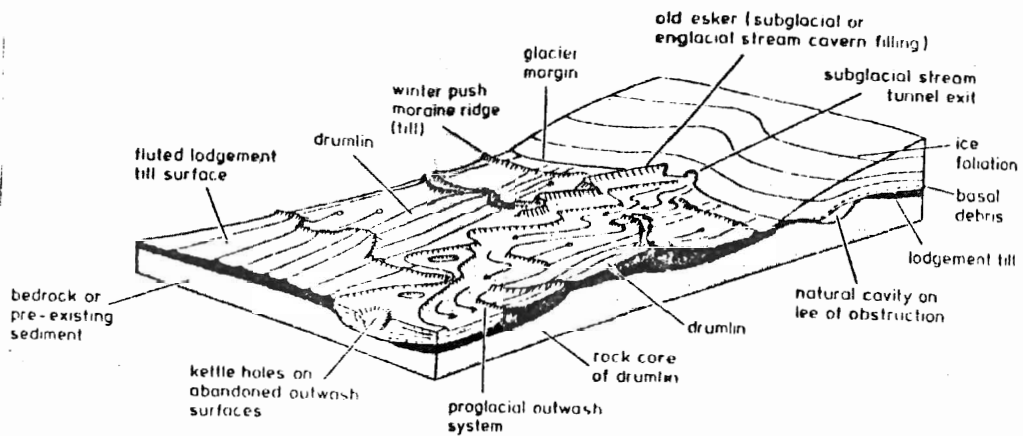
Figure 3.4. Sediment/land form associations.

A) Subglacial/proglacial

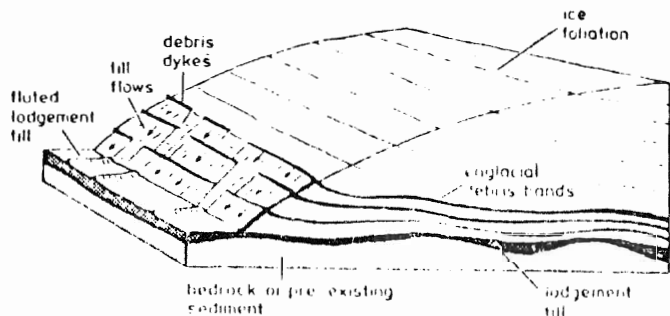
B) Supraglacial

C) Valley

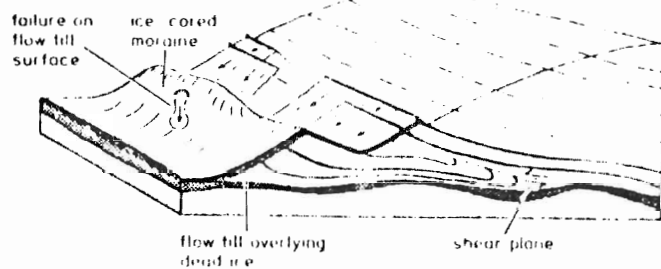
(Boulton and Paul,
1976).



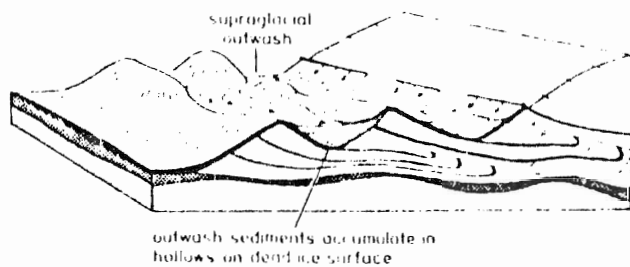
(a)



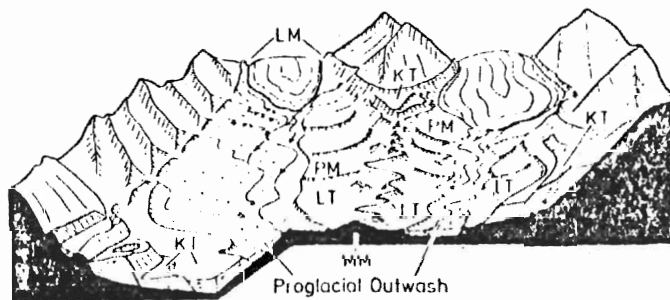
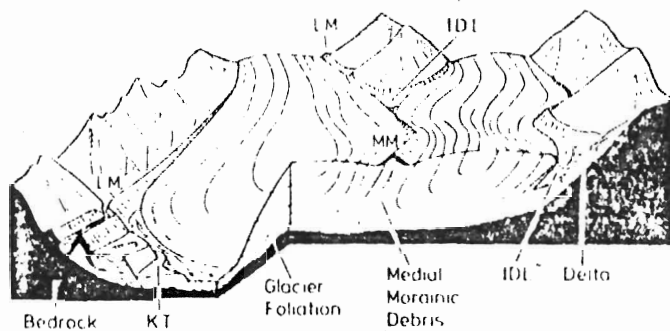
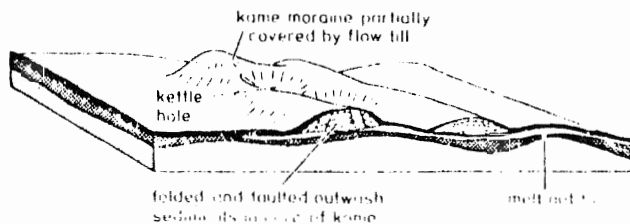
(b)



(c)



(d)



LM - Lateral Moraine
MM - Medial Moraine
IDL - Ice-Dammed Lake

KT - Kame Terrace
LT - Lodgement Till
PM - Push Moraine

and Paul, 1976; Shaw, 1979, Rappol, 1983; Miall, 1983).

When these ideas are compared to unit B sediments at Mohawk Bay, one can see that sedimentation could not have taken place terrestrially. The southern part of the Niagara Peninsula is devoid of any landforms of the type Boulton (1972) and Boulton and Paul (1976) attribute to terrestrial proglacial sedimentation. Furthermore although unit B sediments are somewhat varied, they do not exhibit the characteristics of the sediments associated with kettle holes, outwash fans or eskers. Also the areal extent of these deposits is on the scale of square kilometres (c.f. Cowan, 1975). Therefore proglacial sedimentation in a terrestrial environment is not a likely possibility.

Recently, research into glaciogenic sedimentation in southern Ontario has revealed that glaciolacustrine deposits are more widespread than first thought (c.f. Eyles and Eyles, 1983). These sediments, which show similarities in some cases to glaciomarine sediments, can be sub-divided into proximal and distal regimes.

The proximal zone is considered to be the grounding line of an ice sheet as it transforms into an ice shelf or termination of the ice as a ramp. Grounding line facies varies as a function of seven factors:

- 1) retreat rate of the grounding line
- 2) fluctuations (both long and short term) during retreat

- 3) type of ice front
- 4) water depth
- 5) basal thermal conditions at the grounding line
- 6) ocean (lake) current velocities
- 7) discharge rates of meltwater

It should be stated that these variables apply to both ice shelf and tidewater conditions. Tidewater ice fronts which occur as ramps (c.f. Powell, 1981), are not necessarily restricted to the tidewater glaciomarine environment. This ice marginal configuration, has been attributed to the glaciolacustrine environment as well as ice shelves (Shaw, 1977b; Evenson et.al., 1977).

Proximal zone sediments are usually found comprising morainal banks. These are similar to crossvalley moraines and mark the position of the grounding line of an ice sheet. Sediments are both diamictic and sorted in nature. Diamicts are formed by suspension fallout, gravity flow, bottom melting of an ice shelf and basal squeezing. Sorted and stratified sediments are produced by suspension fallout from over and interflows, underflows, traction currents and low density, low viscosity gravity flows. The latter sediments are mostly sands and gravels since currents tend to winnow and remobilize the silt/clay fraction and transport it basinward. The diamicts are more variable in grain size composition, being more coarse if deposited by gravity flow and finer if deposited by suspension fallout (Powell, 1984).

Distal zone sediments tend to be dominated by laminated and diamict muds, which, in the marine environment, have abundant fauna throughout (Anderson, 1972; Domack, 1982). Sedimentation processes are usually dominated by suspension fallout, iceberg rafting, gravity flows and bottom currents. Laminated and diamict muds are usually the products of the two former processes. Gravity flows in this environment usually take the form of turbidity or contour currents (Wright, et.al., 1983) and result in more sandy deposits although laminated muds can form as a final phase of a turbidite (Walker, 1965). Gravity flow can also be the result of unstable sediment flowing off of topographic highs into troughs in the lake or ocean bottom (Hakanson and Jansson, 1983). Bottom currents also form coarser grained sediment, forming sand or gravel lags, analogous to those in the proximal zone (Powell, 1984).

More common in the glaciolacustrine environment than the glaciomarine, is the rhythmic nature of these sediments. This is more prevalent in the distal sediments where varves are the dominant sediment, reflecting winter sedimentation by suspension fallout and summer sedimentation by turbidity current or underflow activity (Agterberg and Banerjee, 1969). More recently, several authors have shown that localized variations in water depth, rainfall and temperature can result in influxes of coarse sediment which can increase the number of coarse/fine couplets deposited in one year

(Shaw and Archer, 1978; Shaw et al., 1978; Lambert and Hsu, 1979).

In the proximal zone, cyclicity is not as marked. This is mainly due to a greater influx of sediment as well as an increased number of processes operating at the ice margin. Examples of this type of sedimentation are given by Shaw (1977b), Shaw and Archer (1978) and Visser (1983). These workers show a predominance of gravity flow deposits of various types, forming thick sequences of sands and diamicts. Fluctuations of the ice margin can result in deformation of the sediments (Eyles and Eyles, 1984). Deformation in the form of collapse structures may form because of melting of buried ice beneath the proximal sediments (see fig. 14, Shaw, 1977b). Progression of these sediments usually results in the formation of a glaciodeltaic sequence. These deltaic sediments have the usual topset, foreset and bottomset sequences. Bottomset strata are composed of crosslaminated and rippled sands or rhythmically laminated sands or silts and clays. Foreset beds consist of large scale planar crossbeds and are capped by outwash sands and gravels of the topset sediments (Ashley, 1975; Gustavsen et al., 1975; Edwards, 1978; and LeBlanc Smith and Eriksson, 1979). This is illustrated in figure 3.5.

Applying the above comments to the sediments exposed at Mohawk Bay, a glaciolacustrine environment can be spec-

ulated upon for its deposition. Due to the close similarities between Dmm of unit A and B, and the presence of flow structures in Dmm, gravity flow of subglacial debris into the proximal zone is proposed. Powell (1984) states that flowage into the proximal zone of sub or englacial sediments, although changing primary structures, will not change the physical properties substantially. Thus the resedimented debris will have a strong resemblance to its protolith.

Massive flows with clasts suspended within a silt/clay matrix have been described and discussed by Lowe (1979, 1982) and Lawson (1979, 1981, 1982). Lowe (1982) states that the buoyancy and cohesiveness of the silt/clay-water matrix will support the larger clasts during flow, giving rise to a structureless sediment. The resulting deposit is termed a cohesive debris flow or mud flow and may "...include many so-called pebbly mudstones, boulder clays, tilloids, and diamictites..." (p. 293, Lowe, 1982).

The features illustrated by Einsele et.al., (1974), Lawson (1979, 1981, 1982) and Lowe (1979, 1982) are consistent with those found in Dmm, which exhibits substratal rafts of Dmm(r)/Fmd, silt clasts, and push features. Thus it is concluded that Dmm was probably formed by viscous flow, as a cohesive debris flow (Lowe, 1979, 1982; type I Lawson, 1979, 1981, 1982) within a

a proximal glaciolacustrine environment.

It is generally accepted that in a glaciolacustrine or glaciomarine environment, as the distance from the grounding line increases, sediment becomes increasingly finer grained being dominated by muds and diamictic muds (Anderson, 1972; Edwards, 1978). However, Orheim and Elverhoi (1981) and Powell (1984) state that pebbly, fine grained sediments can be found in the proximal zone. This usually occurs under conditions of rapid retreat when subglacial and englacial debris is rafted away quickly. In this situation, diamictic or bergstone muds are found in association with thin, sorted gravel and sand units and other waterlain diamicts. Orheim and Elverhoi (1981) have conducted x-ray examination of these muds from the Weddell Sea and have found the sediment to be devoid of any laminae.

The fine grained, massive nature of Dmm(r)/Fmd and its association with debris flows, suggests a correlation with diamictic muds in the proximal zone. Rafting and the presence of silt clasts are not mentioned by Orheim and Elverhoi (1981) or Powell (1984) but are present in Dmm(r)/Fmd. These were probably due to resedimentation processes. Dmm(r)/Fmd may have initially been deposited on topographic highs. Increased sedimentation would have created instability and resulted in downslope movement of this material. Mass movement in the form of viscous flows

would have caused entrainment of the substrate periodically, in a manner similar to that of Dmm. The lack of turbulence within the flow would have resulted in fine grained particles remaining in the flow and not being remobilized. Had the flow been turbulent, re-mobilization of fines would have caused the formation of a coarse-grained deposit.

The decrease in rafts with increasing stratigraphic height marks a change in depositional process, from a predominance of resedimentation to suspension fallout. As more sediment became deposited, the topographic relief of the depositional surface decreased downslope movement.

As was mentioned earlier, sorted and stratified sediments are quite common in the proximal zone. Powell (1984) states that these can form from suspension fallout, underflow or turbidity currents, or as lags from winnowing processes. With regards to Sm, the latter process can be rejected. Lag deposits which are caused by winnowing would take the form of an inversely graded unit or the coarse unit showing a gradational contact with the underlying unit. In unit B, Sm never exhibits grading and is always in sharp, unconformable contact with respect to the underlying unit.

The two former processes are quite feasible in explaining the occurrence of Sm, and both may be operating simultaneously. Grain size determinations have shown that Sm is quite variable. While pedogenesis may have been the

cause, and probably resulted in alteration from the initial size (c.f. Madgett and Catt, 1977), depositional processes could well have played a part. Suspension fallout would have resulted in the settling of sands, silts and clays into one unit, albeit a graded unit. Underflow and turbidity current mechanisms would form a unit devoid of silts and clays since the turbid nature of the current would mobilize the fines and transport them basinward. Pedogenesis would result in percolation of fines downwards through the unit. Since both Dmm and Dmm(r)/Fmd are fine grained diamicts, they would have a low permeability and serve as a barrier to percolating groundwaters, thus halting downward movement of fines. The silt/clay particles would then be trapped within Sm. This process would remove the primary grading of the unit and result in a massive unit. Redistribution of fines would have effected Sm of underflow or turbidity current origin but since the fines were of low concentration initially the diagenetic change would be limited.

To summarize, it is proposed that unit B was formed in an ice-marginal, glaciolacustrine environment. Dmm was the result of downslope movement of subglacial debris in the form of cohesive debris flows (c.f. Lowe, 1979, 1982). Dmm(r)/Fmd was deposited by two processes, suspension fallout and downslope movements, with the latter decreasing with time, as marked by decreasing amounts of substratal

rafting. Sm was formed by two processes also. Underflows of turbidity currents deposited coarser Sm, devoid of fines, whereas suspension fallout resulted in the deposition of finer grained Sm. Pedogenic processes resulted in the redistribution of fines in Sm, thus altering some of the initial primary characteristics.

3.3 Unit C

As was shown in Chapter 2, unit C is quite different from the other two stratigraphic units. This is illustrated by the presence of the several sand lithofacies, and minor diamict lithofacies. Although deformation has probably extensively altered these sediments from their primary state, it is of a style that is far different from that which effected unit A. Deformation is of water escape variety as opposed to glaciotectonic: this is indicated by the presence of dish-and-pillar structures, flame structures, ball-and-pillows, and convolute laminations. On the basis of these features, a subglacial environment can be rejected. ✓

Soft-sediment deformation in the form of water-escape structures, is common in fine sands to coarse silts which exhibit low cohesion, relatively high permeability and loose packing (Allen, 1971, 1982; Lowe, 1975). One factor that encourages in their deformation is a high sedimentation rate. This causes the underlying sediments

to dewater quickly, resulting in the destruction of primary features. More violent behaviour (dish-and-pillar structures, for example) is caused by the presence of silt/clay layers. These form an impermeable barrier to the upward moving waters, forcing them to move laterally. With increased confining pressure, the silt/clay layer will be punctured, allowing a water/sediment mixture to intrude and deform the overlying layers (Lowe and LoPicollo, 1974; Lowe, 1975; Allen, 1982). This type of behaviour, the upward movement of water/sediment mixtures, is referred to as fluidization (Lowe, 1975; Allen, 1982).

The second type of sediment/water interaction is liquefaction, which involves the upward movement of water with a downward settling of sediment (Lowe, 1975; Allen, 1982). Liquefaction usually results in the mass movement of sediment in the form of debris flows or other gravity mechanisms (Hampton, 1972).

Soft-sediment deformation is not characteristic of a specific environment. Mills (1983) makes the point that the hydrodynamics of a system determine whether or not water release structures can form, and these conditions may be mirrored in several different environments. These structures can be found in alluvial fan, shallow-water marine, fluvial, deltaic, and turbidite environments (Lowe, 1975; Allen, 1982; Mills, 1983).

These environments along with their depositional

Table 3.3

Environments where soft-sediment deformation is common.

Environment	Location	Sedimentary Features
Alluvial fans	Areas of sharp topographic relief. Usually the result of inter-cratonic faulting or uplift.	Predominantly gravel and coarse sands. Comprised of Gm, Gms, Sp, Sh and St. May be associated with braided rivers downstream.
Shallow-water marine	Maritime regions where deposition is dominated by tidal, wave and longshore current processes.	Dependant upon dominant process. Wave current dominated shores are composed of ripple, mega-ripple and sandwave bed forms comprising linear sand ridges. Ridges occur in a series sub-parallel to coast and measure 10m high x 2 to 3 km in width. Tidal-dominated shores are composed of large sandwaves (represented as ridges) showing complex internal structure of dunes, anti-dunes, re-activation surfaces and foreset beds.
Fluvial	Various	Subdivided into straight, braided, meandering and anastomosed. Straight: accumulation of coarse bedload material over short distances. Meandering: long sinuous-rivers characterized by lateral accretion along point bars. Anastomosed: similar to braided except the lack of one predominant channel. Islands tend to be vegetated and aggradation is vertical. Braided: common in glacial systems. Gravel or sand dominated. Sandy braided streams are composed of large bar complexes and sandstone sheets of complex internal structure. Structures found are cross-to plane-bedded sands and dunes.

Environment	Location	Sedimentary features
Deltaic	Mouths of channels entering large bodies of water.	Subdivided into topset, foreset and bottom-set beds. Topset beds approximate braided fluvial conditions. Foresets are composed of large planar crossbeds which are formed by viscous gravity flows. These grade laterally into bottomset beds. Sediments are dominated by mass flows and suspension fallout; may be cyclic in nature.
Turbidites	Various. Resultant of turbid gravity flows travelling along a slope.	Development of Bouma sequence. See figure 3.6.

After: Allen, 1970, 1983.
 Elliot, 1978.
 Gustavsen et.al., 1975.
 Jorgensen, 1982.
 Maddalena, 1982, 1985.
 Miall, 1977, 1978, 1979.
 Rust and Koster, 1984.
 Shaw, 1977b.
 Smith and Putman, 1980.
 Walker, 1965, 1978, 1984.

locations and sedimentary features, are presented in table 3.3. The alluvial fan, shallow-water marine and fluvial environments can all be discounted by both location (for the former two) and sedimentary features. Although water escape structures can be found in these environs, the unit C sediments lack the coarse nature of the alluvial fan and the internal complexities of both shallow-water marine and fluvial conditions.

As discussed earlier in the chapter, deltaic deposits are common in glaciolacustrine situations. Mass flows are the dominant depositional mechanism, which grade laterally from viscous to turbulent varieties. This is followed by a lateral gradation in grain size. Bottomset beds are deposited mainly by turbidites and suspension-fallout (Gustavson, et al., 1975; LeBlanc Smith and Erikson, 1979; Jorgensen, 1982).

Turbidity currents can also be related to morainal banks, as discussed by Powell (1984). These are produced by the outflow of subglacial streams into proglacial lakes (or seas). Two processes result in the formation of these sequences. Bedload sediment may be released at the mouth of the conduit, with finer grained sediments being transported basinward as inter-, or overflows. These sediments would then settle out and undergo reworking by bottom currents (in the form of water/sediment slurries) depositing as turbidites (Powell, 1984; Mackiewicz, et al.,

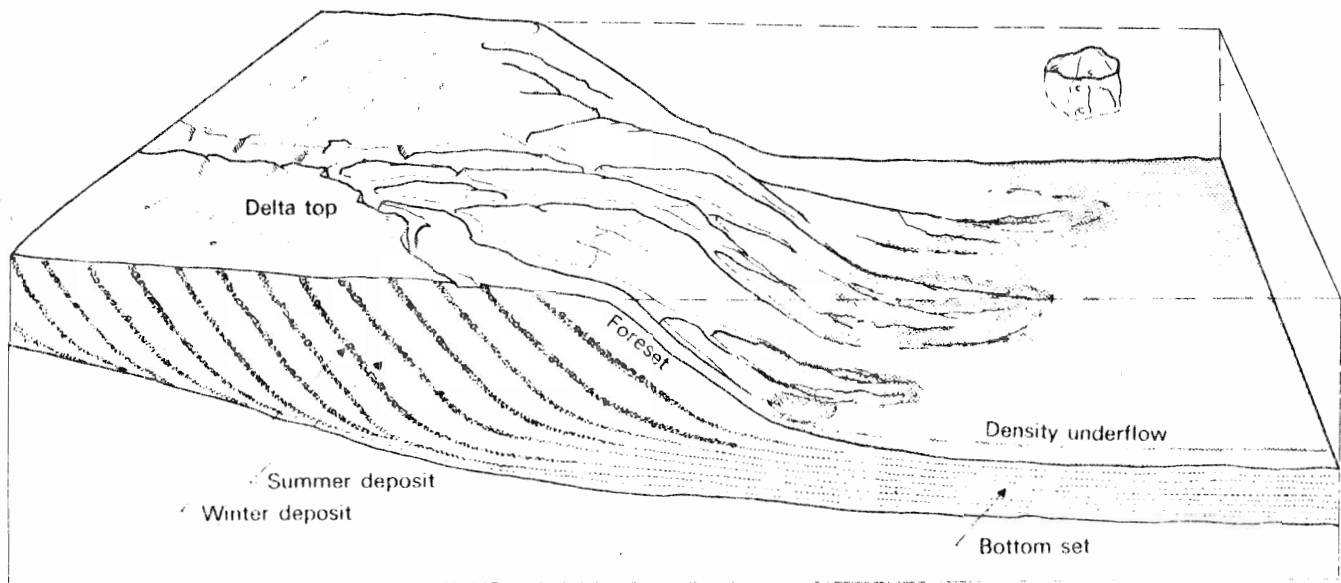


Figure 3.5. Glaciodeltaic sequence. (Edwards, 1978).

1984). High concentrations of sediment within subglacial streams may result in the formation of traction currents. These currents are dense enough to exist as underflows and will result in the deposition of turbidites (Gilbert, 1982; Powell, 1984). Mackiewicz et al., (1984) state that these "distal" products of subglacial streams are deposited within 1.0km of the grounding line.

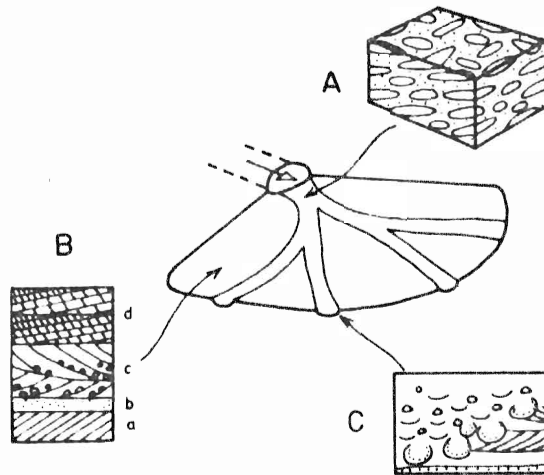
Turbidites have been studied extensively in the non-glacial realm (see Walker, 1984 and references therein). Walker (1978) states that classic turbidites can be distinguished from other gravity flows by "... (1) very parallel bedding, with consistent alternations of sandstone and shale (normally without channelling or major changes in bed thickness laterally): and (2) a consistent set of internal structures that can be described using the Bouma (1962) model." (p 933). The Bouma sequence is shown in figure 3.6.

Work by Lowe (1982) has subdivided turbidites into high and low density varieties. The major difference is that high density turbidity currents have a large percentage of coarse material (pebbles and cobbles) which cannot be carried in suspension by the current and travel as bedload. These form normal to inverse graded beds. Lowe (1982) claims that these deposits, which have been classified as Bouma division A, are not true turbidites, and sediments should not be considered part of the Bouma

Figure 3.7. Sequences within subaqueous outwash fans.

- A) Proximal gravel facies deposited at ice front.
- B) Interchannel facies of (a) large-scale cross-bedded sands, (b) massive sands, (c) trough crossbedded sands, (d) climbing-ripple beds.
- C) distal channel facies of massive, rippled and deformed sands (after Rust, 1977).

Figure 3.6. Bouma sequence (Allen, 1982).



Division	Harms & Fehnestück (1965) Walker (1965, 1967b)
E Pelite	Lower flow regime Plane bed, no movement
D Parallel lamination	Lower flow regime Plane bed, no movement
C Current ripple lamination	Lower flow regime Ripple bed
B Parallel lamination	Upper flow regime Plane bed with grain movement
A Normal grading	Upper flow regime Antidunes

sequence.

Low density turbidites are finer grained sand, silt clay particles that form a series of units with characteristic sedimentary structures (Walker, 1965). This begins with division B, in which upper stage plane beds and dunes are found. Division C is characterized by ripple bedding, commonly climbing ripples of both types A and B, which may grade into convolute laminations. Walker (1978) states that these cross-laminated deposits can consist of either a single or multiple rows of ripples. A single row of climbing ripples indicates reworking of previously deposited sediment by bottom currents whereas multiple rows illustrate active deposition by the turbidity current during rippling. Both Walker (1978) and Allen (1982) claim that division D is very difficult to distinguish from division E. Division D consists of laminated silt and clay, whereas E is mainly formed by pelagic sedimentation. Since the differences in composition are negligible and due to their difficulty in being separated in the field, these two divisions are often combined into division E.

The complete A to E Bouma sequence may not be present. Top-, base-, and middle- absent sequences have been recorded. Top-absent sequences are formed by two mechanisms. Erosion of the upper divisions or increased velocity of the current by the freezing-out of division A. This increased velocity would result in non-deposition of

the upper part of the A division (Walker, 1965).

Base-absent sequences result from late-stage deposition from the turbidity current. Lower divisions may have been deposited nearer the source (Walker, 1965). These base-absent sequences may be the result of flow transformations (Fisher, 1983) from high density, viscous flows (debris or grain flows, for example) to turbulent flows as has been proposed by Hampton (1972), Shaw (1977b) and Lowe (1982). These workers claim that as coarse sediment is deposited from the flow and water contact increases, turbulence becomes the dominant mode of transport. Sequences deposited in this manner will probably consist of BC(D)E or C(D)E divisions.

Walker (1965) states that middle-absent sequences form in a manner similar to top-absent sequences except that mud left in suspension settles upon the basal divisions. Sparks and Wilson (1983) record similar sequences within ash flows from the Aegean Sea.

One can clearly see how soft-sediment deformation can occur within turbidite sequences. Deposition is quite rapid; for division C sediments the work of Ashley et al., (1982), as was mentioned earlier, illustrates that sedimentation rates are on the scale of several centimetres per hour.

Undeformed sediments within unit C at Mohawk Bay show striking similarities to sediments of turbidite affinity.

Those depicted in figures 2.25 to 2.27 show well developed ripples, climbing ripples and minor cross lamination which grade into convolute lamination and, in places, are capped by fine grained or diamict layers. These capping layers are, more often than not, punctured by pillars and are themselves deformed into dish structures. The transitions from Sh to Sr (or St) record B to C division transitions, and the gradation from Sr to Sd may be explained as a transition with division C. Sd/Sm gradations may record either the increasing intensity of dewatering or the incursion of pelagic sedimentation as division (D) E.

The lack of coarse-grained, graded sands and the small extent of Sh, illustrate that these sequences are base-absent sequences. Furthermore the low amount of fine-grained lithofacies indicates that these flows were probably erosive and removed the upper pelagic layers of the substrate. Therefore these sediments are classified as BC, BC(D)E, or C(D)E or thin-bedded turbidites (c.f. Walker, 1978).

The presence of Sr within multiple rows of climbing ripples indicates that these sediments were deposited during the rippling stage of the current. Furthermore, citing the work of Gilbert (1982) and Powell (1984), these turbidity currents probably existed as traction currents, with enough of a sediment load that allowed them to flow along the lake bottom.

The base-absent sequences suggest deposition of coarser-grained material further upstream by viscous flows. Unfortunately, due to lack of exposures, they could not be investigated. It is highly probable that this is the case, since coarse material is dumped very quickly, near the source. Further, these viscous flows probably served to generate the turbulent tractive currents responsible for deposition of exposed sediments.

The presence of Dmm has two ramifications: that the viscous flows did not always transform into turbulent flows and that the ice margin was in close proximity. Dmm in unit C is virtually identical to its counterparts in units A and B, although it has more in common with unit B. Like the latter, it is discontinuous, < 2.0 metres in thickness, has sharp lower (and upper) contacts, silt clasts and truncates underlying laminations. This, along with its association with the thin bedded turbidite sands, indicates deposition by viscous gravity flow. The flow type was probably a cohesion debris flow, utilizing the terminology of Lowe (1979).

That Dmm highly resembles the Dmm of unit A, dictates that the ice margin was close to the site of deposition. Had the ice been a substantial distance away from the site of deposition, the material under flowage would have undergone alterations (due to increasing water content and turbulence) that were sufficient to destroy primary charact-

eristics. Therefore, the presence of slumped subglacial debris strongly suggests proximal sedimentation.

These features, thin bedded turbidites of traction current origin and slumped subglacial debris, along with the extreme local nature of this unit, indicate that the sediment source had to be small scale in relation to the ice sheet as a whole. It is envisaged that a subglacial conduit emanating at the ice margin, subaqueously, was the probable source. Work by Gilbert (1982), Powell (1984) and Mackiewicz et.al., (1984) show that subglacial streams exiting at the ice front, deposit large amounts of varied sediments. These workers have identified the processes of plume sedimentation, traction current activity and debris flow mechanisms to be operating with 0.5 to 1.0 km from the ice front.

Evidence of this style of sedimentation in the Pleistocene has been documented by Rust and Romanelli (1975), Rust (1977) and Cheel and Rust (1982), who have worked in the Ottawa area on Champlain Sea sediments. Termed subaqueous outwash, these sediments are generally made up of gravels and coarse sands with interbedded debris flows near the ice source, which prograde into finer sands and silts, distally. These sands and silts show cross laminations, climbing ripples, and abundant water escape structures. Furthermore, these sediments are found in ridges parallel to ice flow.

Another feature found in the subaqueous outwash near Ottawa, are rather large cut-and-fill channels. These were found to consist of horizontally laminated sands (Rust and Romanelli, 1975; Rust, 1977; Cheel and Rust, 1982). Rust (1977) and Cheel and Rust (1982) attributed these sediments to channelized mass flows which were powerful enough to erode underlying sediments. Flow types were never identified. A diagrammatic representation of this is given in figure 3.7.

Similar cut-and-fill channels were found in unit C at Mohawk Bay. These were composed of massive sands with gravel lags and massive gravels. Like those of Rust and his co-workers, these channels were probably formed by mass flows. Flows may have more dense (less turbulent) than the low density turbidity currents responsible for the surrounding sediments. This is suggested by the presence of the gravel lags and massive nature: higher density flows would possess enough strength to transport coarse material as bedload and would have enough turbulence for the production of sedimentary structures (Walker, 1978; Lowe, 1982).

To summarize, it is postulated that unit C sediments were formed in the distal regions of an ice-marginal, subaqueous outwash which comprised part of a larger morainal bank. This was produced by the outflow from a subglacial conduit. The sand lithofacies were deposited as thin

bedded turbidites from tractive currents which by viscous flows that had undergone flow transformations. These viscous flows (debris or grain flows) probably deposited coarse material nearer the ice source. Rapid sedimentation by the traction currents resulted in widespread soft-sediment deformation. The diamict lithofacies resulted from downslope movements of subglacial debris as cohesive debris flows which did not undergo flow transformations. Cut-and-fill sequences were formed by high density mass flows (turbidites?) depositing massive sands with gravel lags and massive gravels.

Chapter 4

Depositional Model and Regional Implications

4.1 Depositional Model

4.1.1 General

The depositional mechanisms, as discussed in chapter 3, illustrate that the processes involved were widely varied. Hence, the conditions of the ice sheet depositing the sediments must have undergone distinct changes through time. Unit A was deposited subglacially, under active ice conditions. Units B and C, however, were deposited in an ice-proximal position within a proglacial lake. Thus the sequence records the advance and retreat of part of an ice sheet.

Further ramifications are evident. The presence of the sand lithofacies in unit A as meltwater channels show that a warm-melting (c.f. Hughes, 1981) basal thermal regime was in existence. However, since these

channel deposits are preserved, in some places with intact primary structures, and also that localized freezing was occurring. Theoretically, this is what Hughes (1981) claims to be the case in the warm-melting zone: that the ice has a greater percentage of melting than freezing zones, but that freezing does occur.

Ice movement in the area trended northeast to southwest. This is given by fabric, contact lineations and bedrock striations, all of which show parallel orientations. Striae directions were not recorded by the author, however the trend obtained by Feenstra (1982) does correlate directly with the other two directional indicators. The orientation data produced by fabric is a minor problem. Mark (1974) states that if clasts possess a significant plunge, the plunge should be oriented in the up-ice direction. Although Mark (1974) does not state what exactly constitutes a significant plunge, it is suspected that he may mean plunges greater than sub-horizontal ($\leq 10^\circ$). Furthermore, Young (1969) states that plunge does not reflect ice direction. When applied to the data in figure 2.6,

the statements of Young (1969) and Mark (1974) imply that these data are not good directional indicators, due to their variable plunges in both direction and orientation. However, when combined with lineation and striation data, fabric helps support a northeast to southwest trend.

Since the lineations are parallel to ice movement, this suggests that the meltwater channel sediments were transported along a northeast to southwest trend. This is in direct contradiction to the work of Feenstra (1982) who claims these sand inclusions represent proglacial outwash sediment that has undergone ice-push at an ice front parallel to the axis of the ridge. If this were the case, lineations produced by shear would be oriented roughly perpendicular to those measured. Also if these sediments had been produced proglacially, a greater heterogeneity of lithofacies types would be expected, similar to those moraines described by Barnett (1984). These sediments would then have been deformed in a similar manner to those discussed by Thomas (1984) on the Isle of Man. Therefore, this cannot be considered a recessional moraine.

4.1.2

Moraine vs. morainal bank?

The suggestion that units B and C were formed as a morainal bank is fundamentally different from that of a moraine. Although both mark positions of temporary ice front or grounding line stagnation, morainal banks are only confined to the subaqueous regime, whereas moraines are often considered to be of terrestrial origin. The work by Powell (1984) and Mackiewicz et.al., (1984) suggest that deformation of morainal bank sediments due to readvance is of very minor importance. Glacio-tectonic deformation was never recorded in the cores of Mackiewicz et.al., (1984) who sampled sediments to within 0.5km of the ice front. The author suspects that although not reported by these workers, glaciotectionic deformation may occasionally effect morainal bank sediments due to the oscillatory nature of this type of ice/sediment contact. ✓

This type of deformation is recorded by Oldale and

O'Hara (1984), Thomas (1984) and Thomas and Summers (1984) who illustrate the effects of a readvancing ice front, during moraine formation. The work of Thomas (1984) has already been discussed earlier in the thesis. Both Oldale and O'Hara (1984), Thomas and Summers (1984) show development of widespread dislocation and thrusting of proglacial and glacial sediments. In some instances stacking of sediment slices occurs, in a situation analogous to tectonic nappes. Large scale deformation may not always take place. Rains and Shaw (1981) show how moraines in the Antarctic are formed due to progressive meltout of buried, debris-rich ice. In this model the ice front does not experience readvance. Furthermore sediment in this situation is deposited in an arid, terrestrial environment, far different from that of southern Ontario during deglaciation (see Shaw, 1977c, for a more complete discussion of glacial deposition in arid climates).

Although glaciotectonism is in evidence in the sediments at Mohawk Bay, it is restricted only to unit A, which was deposited subglacially. The overlying, morainal bank sediments do not show any glaciotectonic structures.

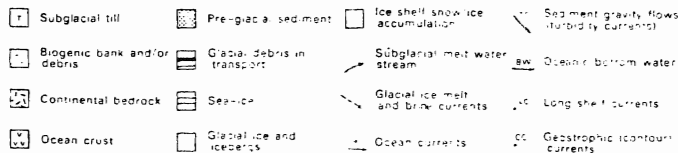
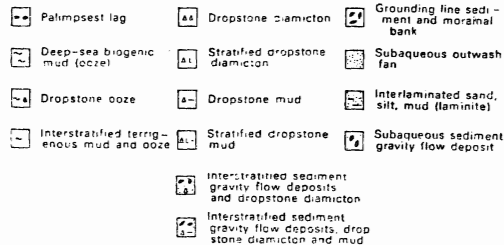
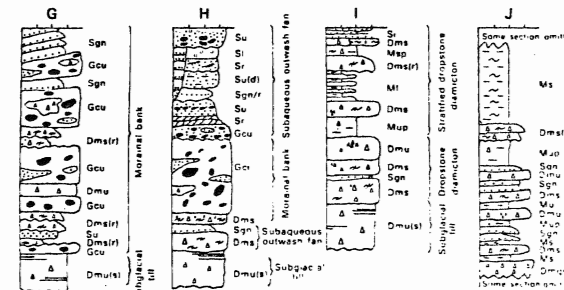
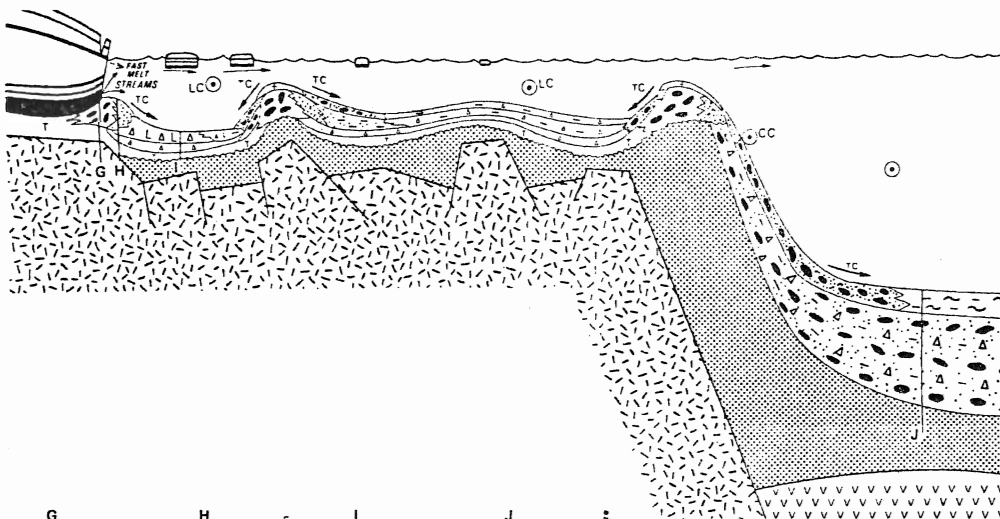
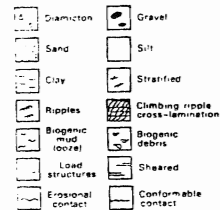
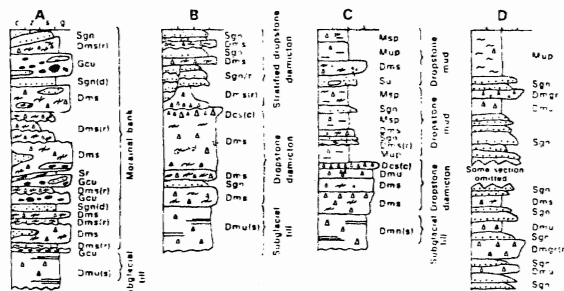
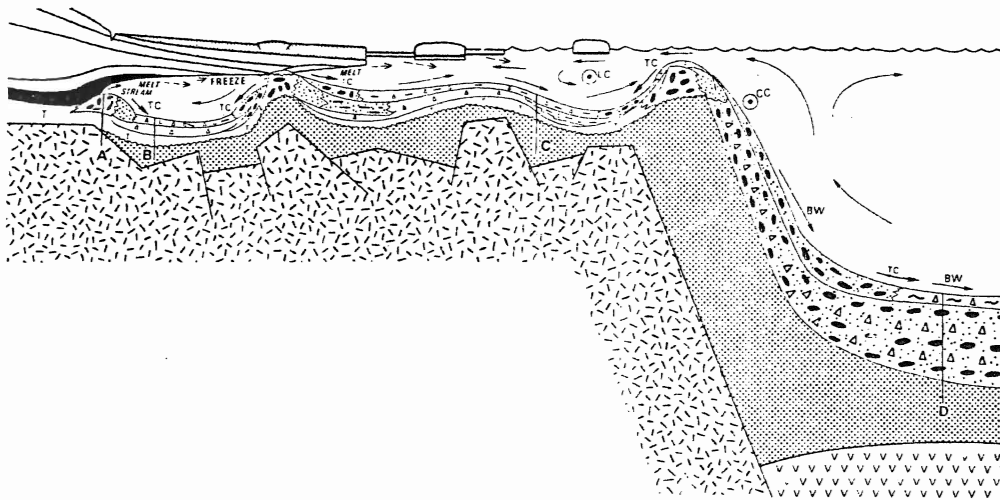
This is consistent with Powell (1984) and Mackiewicz et.al. (1984) in their discussion of morainal banks.

4.1.3

Ice front configuration

As was stated previously, morainal banks can form at the grounding points of an ice shelf or at the base of an ice cliff if a tidewater configuration exists. These situations are depicted in figure 4.1. The vertical profiles illustrate idealized lithofacies associations. What is significant about profiles is that a predominance of coarser grained sediments occurs in the tidewater environment, whereas finer grained, diamictic sediments prevail in the ice shelf environment. These diamictic sediments record deposition by basal meltout and suspension fallout from the ice shelf. Furthermore, due to the occurrence of high velocity underflows, subaqueous outwash fans are more common in the tidewater environs (Gilbert, 1982; Powell, 1984; Mackiewicz et.al.,

Figure 4.1. Idealized ice shelf and tidewater ice front sequences (Powell, 1984).



1984).

Sediments comprising units B and C closely approximate those deposited proximal to a calving ice wall as opposed to beneath an ice shelf. Several factors suggest this. First is that unit C represents sedimentation as a subaqueous outwash fan. Deposition in both units B and C was primarily by high and low density mass flows. $D_{mm}(r)/F_{md}$ in the upper levels of unit B and some S_m beds record the presence of suspension fallout activity: this is only of secondary importance and can be viewed as a subordinate sedimentation process to gravity flows. Although work on Ross Sea Ice Shelf sediments by Anderson (1972), Domack (1982) and Wright *et.al.*, (1983) show that mass flows do occur under ice shelf conditions, subaqueous fans and high density flows are not prevalent.

Secondly, ice shelf sediment would probably be more laterally continuous and regionally extensive. Work by Eyles and Eyles (1983), who attribute sedimentation of the Scarborough Bluffs to be of ice shelf origin, shows that these sediments are deposited "layer-cake" style and are laterally continuous for up to 10km. These

sediments also correlate to those further inland at the Don Valley Brickyards (N. Eyles, personal communication, 1984). Work on modern and ancient ice shelf sediments by Anderson and co-workers also illustrate the areal extent of those deposits.

Last, the relationship between brittle fracture of the ice (in this instance, calving), the tensile stress applied and the yield strength of the ice can be used to determine whether an ice shelf is formed. This relationship is expressed as:

$$\text{brittle fracture} \propto f \left(\frac{\text{tensile stress}}{\text{yield strength}} \right)$$

When an ice sheet decouples from its bed and flows onto water, the friction imparted to the ice by its terrestrial bed is removed, thus increasing tensile stress. In an ice shelf condition, grounding of the ice mass on a topographic high will cause tensile stress to drop to near terrestrial levels, resulting in stabilization of the ice shelf. However, if the ice spreads onto the water body without pinning points, tensile stress will increase dramatically, far exceeding the value of yield

strength, thereby resulting in a calving ice wall.

This is shown to be the case in Antarctica where termination at sea level is marked by a calving ice wall. The only exceptions to this are the Ross, Filchner/Ronne and Amery ice shelves which have been stabilized because of grounding points (Drewry and Robin, 1983).

This situation is further compounded when melting conditions ensue during retreat of the ice sheet. Meltwaters may permeate the ice mass while in retreat. This should result in increased deformation of the ice, such as grain boundary sliding. On the macro-scale, the ice mass may also experience a relative decrease in yield strength since the presence of meltwaters will weaken the ice, compared with an ice sheet that is virtually frozen throughout. Therefore, along with an increased tensile stress, calving rates should increase substantially, further reducing the incidence of shelving.

4.1.4

Deglaciation hypothesis

The above discussion on ice front configurations and calving dynamics leads to an hypothesis for deglaciation of an area with an ice sheet ablating into a proglacial lake basin.

The situation that can be envisaged is the advance of the ice sheet into a proglacial lake as an ice shelf. This is constant throughout the maximum of the lobe's advance. Melting causes tensile stress to increase sharply, inducing brittle fracture, and finally ending in the destruction of the ice shelf and development of a calving ice wall. Due to this release of friction and the presence of meltwater, terrestrial ice velocities should increase and may approach surge conditions. Surging of the ice as an unpinned shelf will result in a spreading of the ice over the water body. This will occur if calving rates cannot keep up with rate of surging. If calving rates approximate surge rates then an ice wall will be preserved. Wave action will help induce fracture and calving in both instances. Such a situation is similar to that of Hughes (1975) in his model for destruction of the present West Antarctic ice

Sheet.

The possibility of surges resulting in the destruction of ice sheet lobes is not unfeasible. Throughout the discussion of depositional processes it has been assumed that the bed of the ice sheet is deformable and mobile, analogous to that of Boulton (1979). This condition is quite different to that favoured by glaciologists of a flat, undeformable bed composed of cubic obstacles of a given dimension (Weertman, 1979; Paterson, 1981). Boulton and Jones (1979) present a model in which the profile and sliding velocity (which governs profile) is controlled by the hydraulic transmissibility of the bed. If the bed allows meltwater to escape through the bed, it will become immobile and behave in a fashion similar to bedrock. Sliding velocities will be slow and profiles will approach a parabola. If the bed does not allow transmission of meltwaters, these waters will collect at the base causing saturation of the bed material. This material will then readily deform. Increasing this meltwater, as in the case of a retreating ice sheet, will increase water pressure at the base of

the ice. This would then result in the decoupling of the ice from its bed and surging will then ensue. This scenario parallels that for sub-polar, surge-type glaciers, as modelled by Clarke et.al., (1984).

The major problem with applying this model to Pleistocene sediments is that a sedimentological expression of the surge should be recorded. In this instance, since increased subglacial meltwaters induce the surge, these waters may also cause melting of the base. With increased velocity, friction will lead to further melting of the ice. Melting of the basal layers will result in deposition of debris. Thus deposition of lodgement till could be expected because of surging.

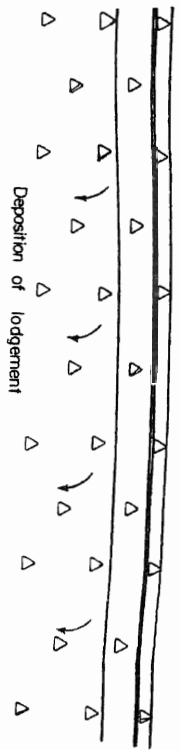
4.1.5

The model

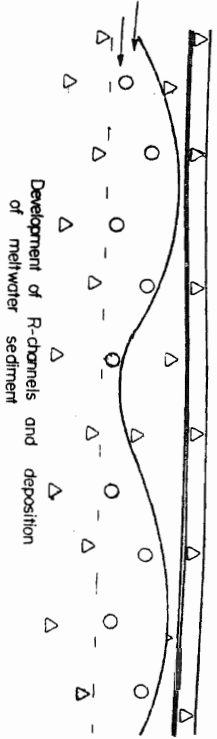
Compilation of the observed characteristics of the sediments, discussion of depositional mechanisms and the implications stated above lead to the development

Figure 4.2. Model for deposition of sediments at
Mohawk Bay.

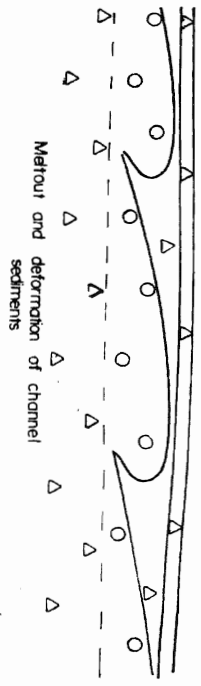
1) Ice movement →



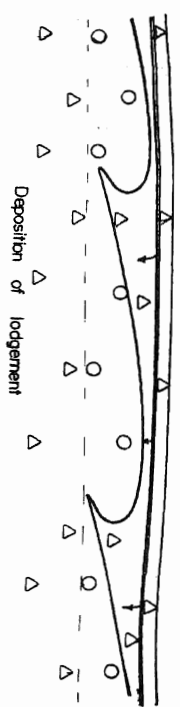
2) _____



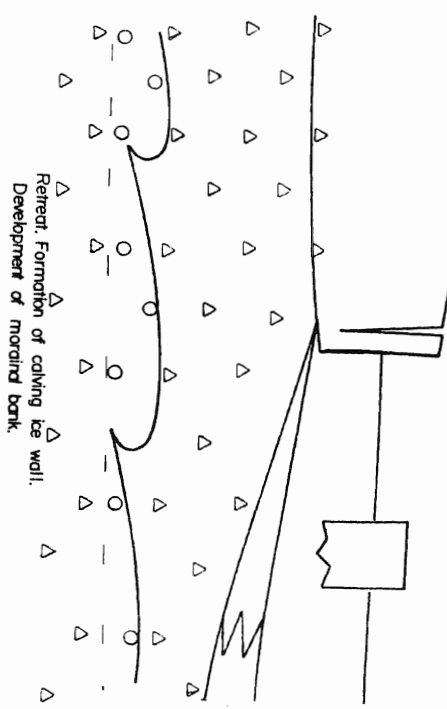
3) _____



4) _____



5) _____



of a model for deposition of the sediments found at Mohawk Bay. This is depicted in figure 4.2

Stage 1 is the initial deposition of lodgement till due to an advancing ice sheet. Within this stage, deformation of pre-glacial sediments, represented by Dmss, occurs due to shearing of the overriding ice.

Stage 2 marks the incursion of meltwaters into the subglacial zone and development of R-channels. These waters were probably sediment-charged slurries which pulsed in a form analogous to turbidity currents. This is indicated by the Sr lithofacies in the form of type A and B climbing ripples (c.f. Walker, 1965). During deposition, the ambient temperatures of the subglacial debris resulted in the migration of the freezing fronts through the saturated meltstream sediments. If the sediments contained diamict layers, their low permeability and porosity would effectively halt the freezing front migration. These unfrozen sediments would then undergo deformation and mixing.

Stagnation of the basal ice due to incorporation of the meltwater sediments initiates stage 3. Stagnation

results from the increased shear strength of the basal ice due to the freezing-on of the sands. Shear stress imparted by the ice cannot overcome the increased shear strength, thus stagnation ensues. This also causes an upperwards shift in the plane of décollement. Continued overriding by the ice will generate deformation (both macro- and meso-scale) within the sand lithofacies. Upon melting, collapse structures in the form of normal faulting will develop from de-watering.

Continued lodgement marks stage 4. It is in this stage that instability due to melting of the ice sheet may occur. What triggers the melting, and ultimately, destruction of the ice sheet, is unknown. However, unpinning of an ice shelf may have occurred resulting in increased velocity of the ice. Increased meltwaters may have caused the development of surge conditions with concurrent lodgement till deposition. As the floating ice continued to breakup, retreat of the ice front resulted. Meltwaters flowing from the ice sheet undoubtedly formed channels.

Stage 5 represents sedimentation in this environ-

ment. Due to a calving ice-wall configuration, with its high energy, underflow mechanisms operating, a morainal bank developed. A subaqueous outwash fan resulted directly from emissions from a subglacial channel. Laterally adjacent to outwash sedimentation, lower energy conditions prevailed and are marked by unit B that is composed of a greater percentage of diamictic sediments than unit C. Unit B sediments were formed mainly by viscous flow and suspension fallout, with turbulent flow occurring occasionally as marked by the coarser grained vacies of Sm. ✓

Stagnation of the ice front, during formation of the morainal bank, was probably short lived. This is suggested by the presence of climbing ripple sequences within the Bouma C unit of preserved turbidite sequences. Although water-escape structures have destroyed the majority of primary structures, the preserved sequences and the amount and intensity of water release all imply that deposition was quite rapid. Indeed, if one applies the rates of deposition of Ashley et.al., (1982) to this sequence, deposition in one melting season

(summer) can be visualized.

Subsequent to stage 5, sedimentation in a proglacial lake occurred.

4.2

Regional implications

The implications of a model such as the one proposed are potentially widespread. Realistically, these can only be suggested since the model developed concerns a small area and [✓]heavily influenced by local conditions. ?
Therefore, only a brief comment on the regional aspects of the model will be offered.

Previously, Feenstra (1982) proposed that the ridge at Mohawk Bay, which he referred to as the Fort Maitland Moraine, formed part of a series of recessional moraines that occur throughout the Niagara Peninsula. These are the Crystal Beach, Fort Erie/Buffalo, Niagara Falls and Vinemount moraines. Feenstra (1982) claims that each record ^S the ice front position as the Ontario-Erie lobe ✓

retreated from the region. If, however the concept of a calving, surge-type glacier is accepted, then the entire deglaciation pattern of the Niagara Peninsula should be questioned.

Detailed sedimentological investigations should be conducted to determine the lithofacies composition, deformation (if any) and relationships with the surrounding sediments. Furthermore, information on ice-front position should be attempted so that a more accurate picture of the deglaciation history can be developed.

Chapter 5

General discussion and conclusions

5.1 General discussion

The model presented in the previous chapter and its associated mechanisms of deposition represents the most feasible way of explaining the sediments exposed at Mohawk Bay. However, while developing the model, a few shortcomings were identified.

Perhaps the most significant shortcoming of the thesis is that all interpretations are based on vertical profiles. Recently, vertical profiling of glacial and fluvial sediments has become popular (e.g. Eyles et al. 1983; Miall, 1977, 1978). The benefits of this approach are that environmental and depositional changes are recorded progressively upwards through a sequence in one specific area. Unfortunately vertical profiles only provide a two-dimensional picture since lateral variations are not taken into account. As mentioned earlier, Allen

(1983) criticized this technique as it applies to fluvial sediment due to the ability of stream to generate quite different sedimentological patterns along its course. This is indeed true of glacial systems. Laterally, glacial sediments can show tremendous variations, which may differ from those vertically.

In an attempt to gain lateral control two avenues were explored: to obtain inland exposures and records of geotechnical borings. Due to the generally flat topography of the southern Niagara Peninsula, good sections through the overburden are not available. What exposures are there are only found along roadside ditches and rarely extend deeper than 1.5m. Therefore these could only be used in a very general way and only with regards to the uppermost sediments.

Geotechnical borings do provide some lateral control, although these are generally restricted to areas of extensive construction. Since the field area is quite rural, borings are almost non existant. Those located further north in Wainfleet and east in Port Colborne give a general view but are too far for correlation purposes.

To obtain a good three-dimensional view, a series of borings inland of the bluff and piston cores offshore within Mohawk Bay should be conducted.

The hypothesis that surging of the Frie-Ontario lobe occurred at the onset of deglaciation may be somewhat contentious. Evenson et al., (1976) and Mickelson et al., (1981) postulate surges for the Lake Michigan lobe, as does Boulton et al., (1977) for the late Devension lobe which flowed along the eastern part of the U.K. In these cases, surging of the ice sheet solved the problem of certain anomalous ¹⁴C dates. In this instance, there are virtually no dateable sediments within the Niagara Peninsula that may support a surge hypothesis.

Furthermore, although Hughes (1975) has proposed surge conditions that would lead to the destruction of the Laurentide Ice Sheet, these ideas and their application to the Laurentide (c.f. Mayewski et al., (1981) have not been accepted by the geologic and glaciologic communities. Thomas (1977) has developed a model for the deglaciation of the St. Lawrence River valley, which does not invoke surging, and has applied this to the present-

day West Antarctic Ice Sheet (c.f. Thomas and Bentley, 1978). The model is as follows. At the glacial maximum, the ice front would exist as an ice shelf. With increased sea levels (for whatever reason), the ice shelf would become ungrounded. Due to the decreased friction, velocity would increase and thinning of the ice mass would occur. This thinning of the ice would lead to decoupling of the ice/bed interface at the grounding line, resulting in its retreat. A new grounding line will be established where thicker ice does not allow decoupling, because of the relation:

$$H_w = 0.9 H_i$$

where H_w represent water depth and H_i ice thickness.

Thomas (1977) states that rates of grounding line retreat will be variable. The initial 150km will be relatively fast, followed by slower retreat of about 300km over 3000 to 6000 years, with a final quick retreat to a position approximately 1100km from the initial grounding line.

This model, although seeming to be successful in explaining the St. Lawrence deglaciation, is not particularly good at explaining deglaciation in the Niagara

Peninsula. Thomas (1977) argues that ice shelf conditions would exist through grounding line retreat. As was stated earlier, sedimentological evidence of shelf conditions is absent at Mohawk Bay. Any other exposures examined along the lake shore only exhibit a diamict that appears to be a correlative of Dmm. Geotechnical borings from Wainfleet and Port Colborne record the presence of laminated silts and clays, diamicts and dropstone diamicts. Due to the nature of these logs, no other interpretations could be attempted. Further, the presence of these sediments are not definite criteria for an ice shelf. Although it does seem that Thomas' (1977) model may not be appropriate, a more regional survey would be necessary to test its applicability.

A further problem of invoking surging just prior to deglaciation exists. Lodgement till may not be deposited by a surging ice mass. As has been well documented (Sugden and John, 1976 and references therein), several modern glaciers are actively depositing lodgement till and are not surging. The problem is though, what sedimentological and geomorphological evidence is left by a surge? There

appear to be two schools of thought: one that states surging conditions result in deposition or that they result in widespread erosion. The latter claims that increased velocity of the ice will increase the abrasion rates along the bed and will lead to extensive scouring. What is not taken into account is the effect of the waters, subglacially. These waters will effectively "buoy-up" the ice mass and should result in reducing abrasion of the bed. Also this assumes that this is moving across a rigid, undeformable substrate. If the bed is incompetent and deformable, the separation between the ice/bed interface may not be as distinct. In this case, increased subglacial meltwaters will saturate the basal debris reducing its frictional resistance. The increased meltwaters should induce further melting of the basal layers leading to melt-out of debris from the ice and deposition of this material. Therefore deposition of lodgement till is quite conceivable with regards to ice with deformable beds, but may not be if the ice rides over a rigid substrate.

Deposition may occur in a different form. If the

basal debris has enough shear strength to withstand the shear stress imparted by the ice, a plane of décollement will be developed above the basal debris layers. This will allow the debris-poor ice to slide over the stagnant debris-rich ice. The resulting sediment formed from the stagnant ice will be a meltout till.

The problem still exists whether the emplacement of lodgement till on top of the meltout facies of unit A represents a surge. This could have been deposited by ice that was travelling at normal velocities. Subsequent to this calving resulted in an ice wall configuration and the scenario was continued as that envisaged in the model. If this was the case, the meltwaters would presumably be transmitted through the basal layer to the subsurface, away from the deforming substrate. The other alternative is that these waters escape before reaching the subglacial zone. The latter is not at all feasible: several workers (e.g Collins, 1979) in order to quantitatively study glacier hydrology, have introduced tracer substances into supraglacial waters, later finding them emitted subglacially. The transmission of the meltwaters to the sub-

surface would only work if the substrate's permeability and porosity were high enough to allow percolation. In this study, Dmm, which would have been the subglacial material, is fine enough not to allow meltwaters to penetrate the subsurface. This is demonstrated presently at Mohawk Bay by observations of groundwater seepage, especially in areas where unit C overlies unit A. After precipitation in the form of rain, or during the spring thaw, waters percolate through the sands comprising unit C. When they come into contact with the underlying Dmm, the waters flow laterally towards the face. This results in large amounts of undercutting and slumping of unit C. Had the permeability and porosity of Dmm been sufficient to allow percolation, lateral movement of groundwaters along the unit A/unit C contact would not occur. Thus, meltwaters would have undoubtedly collected within a basal layer and probably induced a surge.

The meltout subfacies of unit A could be interpreted somewhat differently. In the model proposed, R-channels are formed, meltwater sediments collect, freeze and cause stagnation of the basal layer. Alternatively, N-channels

may have been produced. Meltwaters would then have deposited their sediment load within these depressions and suffered deformation from subsequent lodgement processes without undergoing meltout. This is similar to what Eyles and Sladen (1981) and Eyles et.al. (1982) propose for lodgement tills in Northumberland, England. Conversely, Haldersen and Shaw (1982) and Shaw (1982) claim that the presence of meltwater sediments suggests meltout tills.

The major problem appears to be recognition in the field. Where exposed at modern glaciers, the basal zone shows areas of stagnant ice which would lead to meltout till formation. Such examples are given by Boulton (1971a), Shaw (1977a,c) and Lawson (1979). However, the difficulties arise when dealing with Pleistocene sediments. Structures found in both lodgement and meltout tills can be similar and since both experience subglacial shear, glaciotectionism should be expected in both. Furthermore what is seen as diagnostic features are somewhat subjective and may also be a function of the local area.

With this in mind, the idea that the meltout sub-facies could be lodgement may be possible. However, N-channels may not have been formed. This would be due to the erodability of the ice as opposed to the substrate. Meltwaters flowing subglacially would probably erode the upper debris-poor ice more quickly than the substrate of debris-rich ice. Deposition of the meltstream sediments would undoubtedly be faster than deposition of lodgement till. Freezing of the deposits would produce a hummocky subglacial topography. This would induce stagnation of the ice on both stoss and lee sides of the hummock causing the ice sheet to override the till/meltwater sediment complex. This is similar, but on a greatly reduced scale, to what Shaw (1978)⁹ proposes for the development of Sveg tills and Rogen moraines in Sweden. ?

Finally evidence of proglacial lacustrine sediments (those deposited after stage 5 in the model) is not well defined at Mohawk Bay. In most glaciolacustrine situations, distal sedimentation by the ice is marked by well developed rhythmic or varved sediments (see Shaw, 1977b; Visser, 1983, for examples). Sediments of this type are

not found at all in the exposures. A reconnaissance survey of inland exposures of the uppermost glacial sediments, showed that these sediments were quite similar to Dmm(r)/Fmd. No primary features nor bed contacts relationships could be obtained due to the poor condition of these exposures.

As was stated previously, interlaminated silts and clays, diamicts and dropstone diamicts have been recorded in geotechnical borings at Wainfleet and Port Colborne. Whether they are of similar age or post-date the Mohawk Bay sediments is unknown. Interlaminated silts and clays of this type may not have been deposited at Mohawk Bay because of the relief of the ridge. Since this would have been a topographic high, sediments would have accumulated and suffered flowage into topographic lows (Hakanson and Jansson, 1983). The occurrence of this process is not well developed in the exposures. At lower stratigraphic levels, flow directions appear to be south to southwest. Furthermore, with increasing stratigraphic height, characteristics of flow tend to diminish. Flows may have developed off the ridge, but did not leave

any traceable remnants behind.

Two debatable forms of evidence could be construed as indications of lacustrine activity. First, is the development of interbedded, planar sands and gravels at the extremities of unit C. This is illustrated in figures 2.28 (c) and (d). These are the only occurrence of this type of sedimentation style and are similar to beach foreshore deposits as documented by Reineck and Singh (1980). However, geomorphologic evidence of beach development in the form of wave-cut beaches is absent. Also, such deposits could have been formed by various mechanisms, such as flows of some description.

The second form of evidence could be soft-sediment deformation. Dalrymple (1979) has shown that convolute laminations and ball-and-pillow structures can be formed by wave action. These structures are on the scale of 10.0cm or less. It was further suggested that these structures may indicate a decreasing water depth. Unfortunately at Mohawk Bay, it would be virtually impossible to distinguish wave-generated soft-sediment deformation from non wave-generated types. Therefore,

it is doubtful if this could be applied.

5.2

Conclusions

1) The sediments exposed at Mohawk Bay consist of three stratigraphic units.

2) The basal unit (A) is of subglacial origin, recording a tripartite subfacies association of lodgement-meltout-lodgement processes. The upper and lower lodgement till units are virtually identical. The intervening meltout subfacies consists of subglacial meltwater channel deposits that have been deformed on both macro and small scales from the overriding ice. The upper lodgement subfacies may have been formed during surging of the Ontario-Erie lobe prior to deglaciation.

3) The upper two units (B and C) were formed as a morainal bank during a very brief stagnation (probably one season) of the ice front while in retreat. Unit B, which is only exposed at the flanks of the ridge, is composed

✓

of two diamict lithofacies and a sand lithofacies. These were formed by cohesive debris flow, suspension fallout, and low density (turbidity current?) underflows. Unit C was deposited as a subaqueous outwash fan. Depositional mechanisms were predominantly low density turbidity currents resulting in BC(D)E, C(D)E, BC(D) sequences. Cohesive debris flow and probable high density flows were also operative depositing diamict and channel infill sequences, respectively.

4) The ridge at Mohawk Bay represents a morainal bank deposited at a calving ice wall. This front was oriented roughly perpendicular to the axis of the ridge. An ice wall is favoured rather than ice shelf due to the local nature of the sediments, their relatively coarse grained texture, and the relationship between tensile stress and yield strength of the melting ice sheet to brittle fracture.

REFERENCES:

- AGTERBERG, F.P. and BANERJEE, I., 1969. Stochastic model for the deposition of varves in glacial Lake Barlow-Ojibway, Ontario, Canada: *Canadian Journal of Earth Sciences*, v. 6, p. 625-652.
- ALLEN, J.R.L., 1970. Studies in fluvial sedimentation: A comparison of fining-upwards cyclothems, with special reference to coarse-member composition and interpretation: *Journal of Sedimentary Geology*, v. 40, p. 298-323.
- ALLEN, J.R.L., 1971. A theoretical and experimental study of climbing-ripple cross-lamination, with a field application to the Uppsala esker: *Geografiska Annaler*, v. 53A, p. 157-187.
- ALLEN, J.R.L., 1973. A classification of climbing-ripple cross-lamination: *Journal of the Geological Society of London*, v. 129, p. 537-541.
- ALLEN, J.R.L., 1982. *Sedimentary structures: their character and physical basis*. Volume II: Elsevier, Amsterdam, 663p.
- ALLEN, J.R.L., 1983. Studies in fluvial sedimentation: bars, bar complexes, and sandstone sheets (low-sinuosity braided streams) in the Brownstones (L. Devonian), Welsh Borders: *Sedimentary Geology*, v. 33, p. 237-293.
- ALLEN, J.R.L. and Banks, N.L., 1972. An interpretation and analysis of recumbent-folded deformed cross-bedding: *Sedimentary*, v. 19, p. 257-283.
- ANDERSON, J.B., 1972. Nearshore glacial-marine deposition from modern sediments of the Weddell Sea: *Nature Physical Science*, v. 240, p. 189-192.
- ANDREWS, J.T. and MATSCH, C.L., 1983. *Glacial marine sediments and sedimentation: an annotated bibliography*: Geoabstracts, Norwich, U.K., 227p.
- ASHLEY, G.M., 1975. Rhythmic sedimentation in glacial Lake Hitchcock, Massachusetts-Connecticut: in A.V. Jopling and B.C. McDonald, eds., *Glaciofluvial and glaciolacustrine sedimentation*; Society of Economic Palaeontologists and Mineralogists, Special Publication No. 23, p. 304-320.
- ASHLEY, G.M., SOUTHARD, J.B. and BOOTHROYD, J.C., 1982. Deposition of climbing-ripple beds, a flume simulation: *Sedimentology*, v. 29, p. 67-79.

ASHLEY, G.M., SHAW, J., and SMITH, N.D., 1985. Glacial sedimentary environments; Society of Economic Palaeontologists and Mineralogists, Short Course No. 16, 246p.

BENHAM, P.H., 1975. Glacitectonic structures: a general discussion with particular reference to contorted drift of Norfolk: in, A.E. Wright and F. Moseley, eds.; Ice Ages: Ancient and Modern; Steel House Press, Liverpool, U.K., p. 69-94.

BARNETT, P.J., 1979. Glacial Lake Whittlesey: The probable ice frontal position in the eastern end of the Erie basin: Canadian Journal of Earth Sciences, v. 16, p. 568-574.

BARNETT, P.J., 1984. Glacial stratigraphy and sedimentology central north shore area, Lake Erie, Ontario: Field Trip Guidebook, G.A.C. Annual Meeting, London, Ontario, 41p.

BERTHELTSEN, A., 1978. The methodology of kineto-stratigraphy as applied to glacial geology: Bulletin of the Geological Society of Denmark, v. 27, p. 25-38.

BERTHELTSEN, A., 1979. Recumbent folds and boudinage structures formed by subglacial shear: An example of gravity tectonics: Geologie en Mijnbouw, v. 58, p. 253-260.

BLUEMLE, J.P. and CLAYTON, L., 1984. Large-scale glacial thrusting and related processes in North Dakota: Boreas, v. 13, p. 279-299.

BOLTON, T.E., 1958. Silurian stratigraphy and palaeontology of the Niagara Escarpment in Ontario: Geological Survey of Canada, Memoir 289, 145p.

BOULTON, G.S., 1968. Flow tills and related deposits on some Vestspitzbergen glaciers: Journal of Glaciology, v. 7, p. 391-412.

BOULTON, G.S., 1970a. On the origin and transport of englacial debris in Svalbard glaciers: Journal of Glaciology, v. 9, p. 213-219.

BOULTON, G.S., 1970b. On the deposition of subglacial and meltout tills at the margins of certain Svalbard glaciers: Journal of Glaciology, v. 9, p. 231-245.

BOULTON, G.S., 1972a. Modern arctic glaciers as depositional models for former ice sheets: Journal of the Geological Society of London, v. 128, p. 361-393.

BOULTON, G.S., 1975. Processes and patterns of subglacial sedimentation: In A.E. Wright and F. Moseley, eds.; *Ice Ages: Ancient and Modern*, Steel House Press, Liverpool, U.K. p. 7-42.

BOULTON, G.S., 1979. Processes of glacier erosion on different substrata: *Journal of Glaciology*, v. 23, p. 15-37.

BOULTON, G.S. and JONES, A.S., 1979. Stability of temperate ice caps and ice sheets resting on beds of deformable sediment: *Journal of Glaciology*, v. 24, p. 29-43.

BOULTON, G.S. and PAUL, M.A., 1976. The influence of genetic processes on some geotechnical properties of glacial tills: *Quarterly Journal of Engineering Geology*, v. 9, p. 159-194.

BOULTON, G.S., DENT, D.L. and MORRIS, E.M., 1974. Subglacial shearing and crushing and the role of water pressure in tills from southeast Iceland: *Geografiska Annaler*, v. 56A, p. 135-145.

BOULTON, G.S., JONES, A.S., CLAYTON, K.M. and KENNING, M.J., 1977. A British ice-sheet model and patterns of glacial erosion and deposition in Britain: in *British Quaternary Studies, Recent Advances*; F.W. Shotton, ed.; Clarendon Press, Oxford, p. 231-246.

CHAPMAN, L.J. and PUTNAM, D.F., 1966. The physiography of southern Ontario, 2nd. edition: University of Toronto Press, Toronto.

CHEEL, R.J. and RUST, B.R., 1982. Coarse-grained facies of glaciomarine deposits near Ottawa, Canada: in *Research in glacial, glaciofluvial and glaciolacustrine systems*; R. Davidson-Arnott, W. Nickling and B.D. Fahey, eds.; Geoabstracts, Norwich, U.K., p. 279-295.

CLARKE, G.K.C., COLLINS, S.G. and THOMPSON, D.E., 1984. Flow, thermal structure and subglacial conditions of a surge-type glacier: *Canadian Journal of Earth Sciences*, v. 21, p. 232-240.

COLLINS, D.N., 1979. Quantitative determination of the subglacial hydrology of two alpine glaciers: *Journal of Glaciology*, v. 23, p. 347-362.

COWAN, W.R., 1976. Quaternary geology of the Orangeville area, southern Ontario: Ontario Division of Mines, *Geoscience Report* 141, 98p.

DALRYMPLE, R.W., 1979. Wave-induced liquefaction: a modern example from the Bay of Fundy: *Sedimentology*, v. 26, p. 835-844.

DREIMANIS, A., 1977. Correlation of Wisconsin glacial events between the eastern Great Lakes and the St. Lawrence lowlands: *Geographic Physique et Quaternaire*, v. 31, p. 37-51.

DREIMANIS, A., 1979. The problems of waterlain tills: in *Moraines and Varves*, Ch. Schluter, ed.; A.A. Balkema, Rotterdam, p. 167-177.

DREWRY, D.J. and ROBIN, G. deQ., 1983. Form and flow of the Antarctic Ice Sheet during the last million years: in, *The climatic record in polar ice sheets*; G. deQ. Robin, ed.; Cambridge University Press, Cambridge, p. 28-38.

EDWARDS, M.B., 1978. Glacial environments: in H.G. Reading, ed.; *Sedimentary Environments and Facies*; Blackwell Scientific Publications; London, p. 416-438.

EHLERS, J. and STEPHEN, H.J., 1979. Forms at the base of till strata as indicators of ice movement: *Journal of Glaciology*, v. 22, p. 345-355.

EINSELE, G., OVERBECK, R., SCHWARZ, H. U. and UNSOLD, G., 1974. Mass physical properties, sliding and erodability of experimentally deposited and differentially consolidated clayey muds: *Sedimentology*, v.21, p. 339-372.

EVENSON, E.B., FARRAND, W.R., ESCHMAN, D.F. MICKELSON, D.M. and MAHER, J., 1976. Greatlakean substage: a replacement for Valderan substage in the Lake Michigan basin: *Quaternary Research*, v.6, p.411-424.

EVENSON, E.B., DREIMANIS, A. and NEWSOME, J.W., 1977. Subaquatic flow tills: a new interpretation for the genesis of some laminated till deposits: *Boreas*, v.6, p. 115-133.

EYLES, C.H. and EYLES, N., 1983. Sedimentation in a large lake: a re-interpretation of the late Pleistocene stratigraphy at Scarborough Bluffs, Ontario, Canada: *Geology*, v. 11, p. 146-152.

EYLES, C.H. and EYLES, N., 1984. Glaciomarine sediments of the Isle of Man as a key to late Pleistocene stratigraphic investigations in the Irish Sea Basin: *Geology*, v. 12, p. 359-364.

EYLES, N. and MIALL, A.D., 1984. Glacial facies: in R.G. Walker, ed.; *Facies Models*, 2cd. edition, Geoscience Canada, Reprint Series 1, p. 15-38.

EYLES, N. and SLADEN, J.A., 1984. Stratigraphy and geotechnical properties of weathered lodgement till in Northumberland, England: *Quarterly Journal of Engineering Geology*, v. 14, p. 129-181.

EYLES, N., SLADEN, J.A. and GILROY, S., 1982. A depositional model for stratigraphic complexes and facies superimposition in lodgement tills: *Boreas*, v.11, p. 317-333.

EYLES, N., EYLES, C.H. and MIALL, A.D., 1983. Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict and diamictite sequences: *Sedimentology*, v. 30, p. 393-410.

FEENSTRA, B.H., 1974. Quaternary geology of the Dunnville area: Ontario Division of Mines, Preliminary Map. p. 981.

FEENSTRA, B.H., 1981. Quaternary geology and industrial minerals of the Niagara-Welland area, southern Ontario: Ontario Geological Survey, Open File Report 5361, 260p.

FORDHAM, C.J., 1981. On the origins of the Port Maitland Moraine: Unpublished BSc thesis, Brock University, St. Catharines, Ontario.

FULLERTON, D.S., 1980. Preliminary correlation of post-Erie Interstadial events (16000-10000 radiocarbon years beforepresent), central and eastern Great Lakes region, and Hudson and Champlain, and St. Lawrence Lowlands, United States and Canada: United States Geological Survey, Professional Paper 1089, 52.

GIBBOARD, P., 1980. The origin of stratified Catfish Creek Till by basal melting: *Boreas*, v.9, p. 71-85.

GILBERT, R., 1982. Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: Glaciomarine processes in fjords of eastern Cumberland Peninsula: *Arctic and Alpine Research*, v. 14, p. 1-12.

GROVENOR, C.P., 1980. Heavy minerals and sedimentological studies on the glacial Late Precambrian Gaskiers Formation of Newfoundland: *Canadian Journal of Earth Sciences*, v. 17, p. 1331-1341.

- GROVENOR, C.P., von BRUNN, V. and DREIMANIS, A., 1984. Nature and classification of waterlain glaciogenic sediments, exemplified by Pleistocene, Late Palaeozoic and Late Pre-Cambrian deposits: *Earth Science Reviews*, v. 20, p. 105-166.
- GUSTAVSON, T.C., ASHLEY, G.M. and BOOTHROYD, J.C., 1975. Depositional sequences in glaciolacustrine deltas: in Jopling, A.V. and McDonald, B.C., eds.; *Glaciofluvial and glaciolacustrine Sedimentation*, Society of Economic Palaeontologists and Mineralogists Special Publication 23, p. 264-280.
- HAKANSON, L. and JANSSON, M., 1983. Principles of lake sedimentology; Springer-Verlag, New York, 316p.
- HALDERSEN, S. and SHAW, J., 1982. The problem of recognizing meltout till: *Boreas*, v. 11, p. 261-277.
- HAMPTON, M.A., 1972. The role of subaqueous debris flow in generating turbidity currents: *Journal of Sedimentary Petrology*, v. 42, p. 775-793.
- HARRIS, C. and BOTHAMLEY, K., 1984. Englacial deltaic sediments as evidence for basal freezing and marginal shearing, Leirbreen, Southern Norway: *Journal of Glaciology*, v. 30, p. 30-34.
- HOBBS, B.E., MEANS, W.D. and WOLTHAMS, P.F., 1976. An outline of structural geology; John Wiley and Sons, New York, 571p.
- HUGHES, T., 1975. The West Antarctic Ice Sheet: instability disintegration and initiation of ice ages: *Reviews of Geophysics and Space Physics*, v. 13, p. 502-526.
- HUGHES, T.J., 1981. Numerical reconstruction of palaeo-ice sheets: in *The Last Great Ice Sheets*; G.H. Denton and T.J. Hughes, eds.; John Wiley and Sons, New York; p. 221-261.
- JORGENSEN, N.B., 1982. Turbidites and associated re-sedimented deposits from a tilted glaciodeltaic sequence, Denmark: *Dann. geol. Unders. Arbog* 1981, p. 47-72.
- KARROW, P.F., 1984. Quaternary stratigraphy and history, Great Lakes-St. Lawrence region: in *Quaternary Stratigraphy of Canada - A Canadian contribution to IGCP project 24*; R.J. Fulton, ed.; Geological Survey of Canada, Paper 84-10, p. 137-153.

KARROW, P.F. and TERASMAE, J., 1970. Pollen-bearing sediments of the St. Davids buried valley fill at the Whirlpool, Niagara River gorge, Ontario: Canadian Journal of Earth Sciences, v. 7, p. 539-541.

KRUGER, J., 1979. Structures and textures in till indicating subglacial lodgement: Boreas, v. 8, p. 323-340.

LAMBERT, A. and HSU, K.J., 1979. Non-annual cycles of varve-like sedimentation in Walmsee, Switzerland: Sedimentology, v. 26, p. 453-461.

LAWSON, D.E., 1979. Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska: Cold Regions Research and Engineering Laboratory Report 79-9, 112p.

LAWSON, D.E., 1981. Sedimentological characteristics and classification of depositional processes and deposits in the glacial environment: Cold Regions Research and Engineering Laboratory Report 81-27, 16p.

LAWSON, D.E., 1982. Mobilization, movement, and deposition of active subaerial sediment flows, Matanuska Glacier, Alaska: Journal of Geology, v. 90, p. 279-300.

LeBLANC SMITH, G. and ERIKSON, K.A., 1979. A fluvio-glacial and glaciolacustrine deltaic depositional model for Permo-Carboniferous coals of the northeastern Karoo Basin, South Africa: Palaeogeography, Palaeo-climatology, Palaeoecology, v. 27, p. 67-84.

LOWE, D.R., 1975. Water escape structures in coarse-grained sediments: Sedimentology, v. 22, p. 157-204.

LOWE, D.R., 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits: in Geology of Continental Slopes, ed. L.J. Doyle and O.H. Pilkey, Society of Economic Palaeontologists and Mineralogists Special Publication 27, p. 61-73.

LOWE, D.R., 1982. Sediment gravity flows: II. Depositional models with special reference to the high-density turbidity currents: Journal of Sedimentary Petrology, v. 52, p. 270-297.

LOWE, D.R. and LOPICOLA, R.D., 1974. The characteristics and origins of dish and pillar structures: Journal of Sedimentary Petrology, v. 44, p. 484-501.

MADDALENA, A.L., 1982. A study of stream selective sorting and experimental abrasion applied to Twenty Mile Creek, Jordan, Ontario: unpublished BSc thesis, Brock University, St. Catharines, Ontario. 102p.

MADDALENA, A.L., 1984. Armour coats inverse grading, and streambed scour in selected streams of southern Ontario and western New York: unpublished MSc thesis, Brock University, St. Catharines, Ontario, 154p.

MADGETT, P.A. and CATT, J.A., 1978. Petrography, stratigraphy and weathering of late Pleistocene tills in East Yorkshire, Lincolnshire and North Norfolk: Proceedings of the Yorkshire Geological Society, v. 42, p. 55-108.

MARK, D.M., 1974. On the interpretation of till fabrics: Geology, v.2, p. 101-104.

MAYEWSKI, P.A., DENTON, G.H. and HUGHES, T.J., 1981. Late Wisconsin ice sheets of North America: in The Last Great Ice Sheets; G.H. Denton and T.J. Hughes, eds.; John Wiley and Sons, New York, p. 67-178.

MCCABE, A.M., DARDIS, G.F. and HARVEY, P.M., 1984. Sedimentology of a late Pleistocene submarine-moraine complex, County Down, Northern Ireland: Journal of Sedimentary Petrology, v. 54, p. 716-730.

MENZIES, J. 1981a. Freezing fronts and their possible influence upon processes of subglacial erosion and deposition: Annals of Glaciology, v. 2, p. 52-56.

MENZIES, J., 1981b. Temperatures within subglacial debris - A gap in our knowledge: Geology, v.9, p. 271-273.

MENZIES, J., 1984. Drumlins: an annotated bibliography, geoabstracts, Norwich, U.K.

MIALL, A.D., 1977. A review of the braided river depositional environment: Earth Science Reviews, v. 13, p. 1-62.

MIALL, A.D., 1978. Lithofacies types and vertical profile models in braided river deposits: a summary: in A.D. Miall, ed., Fluvial Sedimentology; Canadian Society of Petroleum Geologists; Memoir 5, p. 597-604.

MIALL, A.D., 1979. Mesozoic and Tertiary geology of Banks Island, Arctic Canada. The history of an unstable croton: Geological Survey of Canada, Memoir 387, 235p.

MIALL, A.D., 1983. Glaciofluvial transport and deposition: in N. Eyles, ed.; Glacial Geology, An introduction for engineers and earth scientists; Pergamon Press; Oxford, U.K.; p. 168-183.

MICKELSON, D.M., ACOMB, L.J. and BENTLEY, C.R., 1981. Possible mechanism for the rapid advance and retreat of the Lake Michigan lobe between 13000 and 11000 years BP: Annals of Glaciology, v. 2, p. 185-186.

MILLS, P.C., 1983. Genesis and diagnostic value of soft-sediment deformation structures - A review: Sedimentary Geology, v. 35, p. 83-104.

MORGENSTERN, N.R., ROGGENSACK, W.D. and WEAVER, J.S., 1980. The behaviour of friction piles in ice and ice-rich soils: Canadian geotechnical Journal, v. 17, p. 405-415.

MULLER, E.H., 1977. Quaternary geology of New York, Niagara Sheet: New York State Museum and Science Service, Map and chart series, no. 28.

NARDIN, T.R., HEIN, F.J., GORSHINE, D.S. and EDWARDS, B.D., 1979. A review of mass movement processes, sediment and acoustic characteristics and contrasts in slope and base-of-slope systems versus canyon-fan-basin floor systems: in Geology of Continental Slopes, eds L.T. Doyle and O.H. Pilkey, Society Economic Palaeontologists and Mineralogists special publication 27, p. 61-73.

NIXON, J.F. and McROBERTS, E.C., 1976. A design approach for pile foundations in permafrost: Canadian Geotechnical Journal, v. 13, p. 40-57.

OLDALE, R.N. and O'HARA, C.J., 1984. Glaciotectonic origin of the Massachusetts coastal end moraines and a fluctuating late Wisconsinan ice margin: Geological Society of America Bulletin, v. 95, p. 61-74.

ORHEIM, O. and ELVERHOI, A., 1981. Model for submarine glacial deposition: Annals of Glaciology, v. 2, p. 123-128.

PATERSON, W.J.B., 1981. The physics of glaciers, 2nd edition: Pergamon Press; Oxford, U.K.; 380p.

POWELL, R.D., 1981. A model for sedimentation by tidewater glaciers: Annals of Glaciology, v. 2, p. 129-134.

POWELL, R.D., 1984. Glacimarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples: Marine Geology, v. 57, p. 1-52.

- PREST, V.K., 1969. Quaternary geology : in *Geology and economic minerals of Canada*, R.J.W. Douglas, ed.; p. 675-764.
- RAINS, R.B. and SHAW, J., 1981. Some mechanisms of controlled moraine development, Antarctica: *Journal of Glaciology*, v. 27, p. 113-128.
- RAPPOL, M., 1983. Glacigenic properties of till: *Studies in glacial sedimentology from the Allgau Alps and the Netherlands*: Unpublished PhD Dissertion, University of Amsterdam, 225p.
- REMECK, H.E. and SINGH, I.B., 1980. *Depositional sedimentary environments*; Springer-Verlag, New York, 549p.
- RICKARD, L.V. and FISHER, D.W., 1970. Geologic map of New York, Niagara Sheet: New York State Museum and Science Service, Map and Chart Series, No. 15.
- ROGGENSACK, W.D. and MORGENSTERN, N.R., 1978. Direct shear tests on natural fine-grained permafrost soils: *Proceedings Third International Conference on Permafrost*; Edmonton, Alberta, Canada, V. 1, p. 728-735.
- RUST, B.R., 1977. Mass flow deposits in a Quaternary succession near Ottawa, Canada: Diagnostic criteria for subaqueous outwash: *Canadian Journal of Earth Sciences*, v. 14, p. 175-184.
- RUST, B.R. and KOSTER, E.H., 1984. Coarse alluvial deposits: in R.G. Walker ed.; *Facies Models*, 2cd. edition, Geoscience Canada, Reprint Series, p. 53-70.
- RUST, B.R. and ROMANELLI, R., 1975. Late Quaternary subaqueous outwash deposits near Ottawa, Canada: in A.V. Jopling and B.C. McDonald, eds.; *Glaciofluvial and glaciolacustrine sedimentation*; Society of Economic Palaeontologists and Mineralogists, Special Publication No. 23, p.177-192.
- SHAW, J., 1977a. Till body morphology and structure related to glacier flow: *Boreas*, v. 6, p. 189-201.
- SHAW, J., 1977b. Sedimentation in alpine lake during deglaciation, Okanagan Valley, British Columbia, Canada: *Geografiska Annaler*, v. 594, p. 221-240.
- SHAW, J., 1977c. Tills deposited in arid environments: *Canadian Journal of Earth Sciences*, v. 14, p. 1239-1245.

- SHAW, J., 1979. Genesis of the Sveg tills and Rogen moarines of central Sweden: a model of basal meltout: *Boreas*, v. 8, p. 409-426.
- SHAW, J., 1982. Meltout till in the Edmonton area, Alberta, Canada: *Canadian Journal of Earth Sciences*, v. 19, p. 1548-1569.
- SHAW, J., 1983. Drumlin formation related to inverted melt-water erosional marks: *Journal of Glaciology*, v. 29, p. 461-479.
- SHAW, J. and ARCHER, J., 1978. Winter turbidity current deposits in late Pleistocene glaciolacustrine varves, Okanagan Valley, British Columbia, Canada: *Boreas*, v. 7, p. 123-130.
- SHAW, J., GILBERT, R. and ARCHER, J.J.J., 1978. Pro-glacial lacustrine sedimentation during winter: *Arctic and Alpine Research*, v. 10, p. 689-699.
- SMITH, D.G. and PUTNAM, P.E., 1980. Anastomosed river deposits: modern and ancient examples in Alberta, Canada: *Canadian Journal of Earth Sciences*, v. 17, p. 1396-1406.
- SPARKS, R.S.J. and WILSON, C.J.N., 1983. Flow-head deposits in ash turbidites: *Geology*, v. 11, p. 348-351.
- ST. JACQUES, D.A. and RUKAVIVA, N.A., 1973. Lake Erie nearshore sediments: Mohawk Point to Point Burwell, Ontario: *Proceedings 16th. Conference of Great Lakes Research; International Association of Great Lakes Research*, p. 454-467.
- SUGDEN, D.E. and JOHN, B.S., 1976. *Glaciers and landscapes*; Edward Arnold Publishers, London, 376 p.
- TELFORD, P.G. and TARRANT, G.A., 1975. Palaeozoic geology of the Welland-Fort Erie Area, southern Ontario: Ontario Division of Mines, Preliminary Map, p. 989.
- TERASMAE, J., 1979. Radiocarbon dating and polynology of glacial Lake Nipissing deposits at Wasaga Beach, Ontario: *Journal of Great Lakes Research*, v. 5, p. 292-300.
- TERASMAE, J., 1980. Some problems of late Wisconsin history and geochronology in southeastern Ontario: *Canadian Journal of Earth Sciences*, v. 17, p. 361-381.
- TERASMAE, J., KARROW, P.F. and DREIMANIS, A., 1972. Quaternary stratigraphy and geomorphology of the eastern Great Lakes region of southern Ontario: *International*

Geological Congress, 24th. session, excursion A42, Montreal, 75p.

TERASMAE, J. and MATTHEWS, H.L., 1980. Late Wisconsin white spruce (Picea glauca (Moench) Voss) at Brampton, Ontario: Canadian Journal of Earth Sciences, v. 17, p. 1087-1095.

THOMAS, G.S.P., 1984. The origin of the glacio-dynamic structure of the Bride Moraine, Isle of Man: Boreas, v. 13, p. 355-364.

THOMAS, G.S.P. and SUMMERS, A.J., 1984. Glacio-dynamic structures from the Blackwater Formation, County Wexford, Ireland: Boreas, v. 13, p. 5-12.

THOMAS, R.H., 1977. Calving bay dynamics and ice sheet retreat up the St. Lawrence valley system: Geographic Physique et Quaternaire, v. 31, p. 347-356.

THOMAS, R.H. and BENTLEY, C.R., 1978. A model for Holocene retreat of the West Antarctic Ice Sheet: Quaternary Research, v. 19, p. 150-170.

VISSER, J.N.J., 1983. A lacustrine glacier retreat sequence from the Permo-Carboniferous Dwyka Formation, Republic of South Africa: Journal of Glaciology, v. 29, p. 515-520.

WALDER, J.S., 1982. Stability of sheet flow of water beneath temperate glaciers and implications for glacier surging: Journal of Glaciology, v. 28, p. 273-293.

WALKER, R.G., 1978. Deep-water sandstone facies and ancient submarine fans: Models for exploration for stratigraphic traps: American Association of Petroleum Geologists Bulletin, v. 62, p. 932-966.

WALKER, R.G., 1984. Shelf and shallow marine sands: in R.G. Walker, ed., Facies Models, 2nd. edition, Geoscience Canada, Reprint Series 1, p. 141-170.

WALKER, R.G. and CANT, D.J., 1984. Sandy fluvial systems: in R.G. Walker, ed., Facies Models, 2nd. edition, Geoscience Canada, Reprint Series 1, p. 71-90.

WEERTMAN, J. and BIRCHFIELD, G., 1983. Stability of sheet water under a glacier: Journal of Glaciology, v. 29, p. 374-382.

WINN, C.E., 1978. Vegetation history and geochronology of several sites in south-southwestern Ontario with discussion on mastodon extinction in southern Ontario: unpublished MSc thesis, Brock University, St. Catharines, Ontario, 374p.

WRIGHT, R., ANDERSON, J.B. and FISCO, P.P., 1983. Distribution and association of sediment gravity flow deposits and glacial/glacial-marine sediments around the continental margin of Antarctica: in Glacial-marine Sedimentation, B.F. Molnia, ed., Plenum Press, New York, p. 265-300.

YOUNG, J.A.T., 1969. Variations in till macrofabric over very short distances: Geological Society of America bulletin, v. 80, p. 2343-2352.

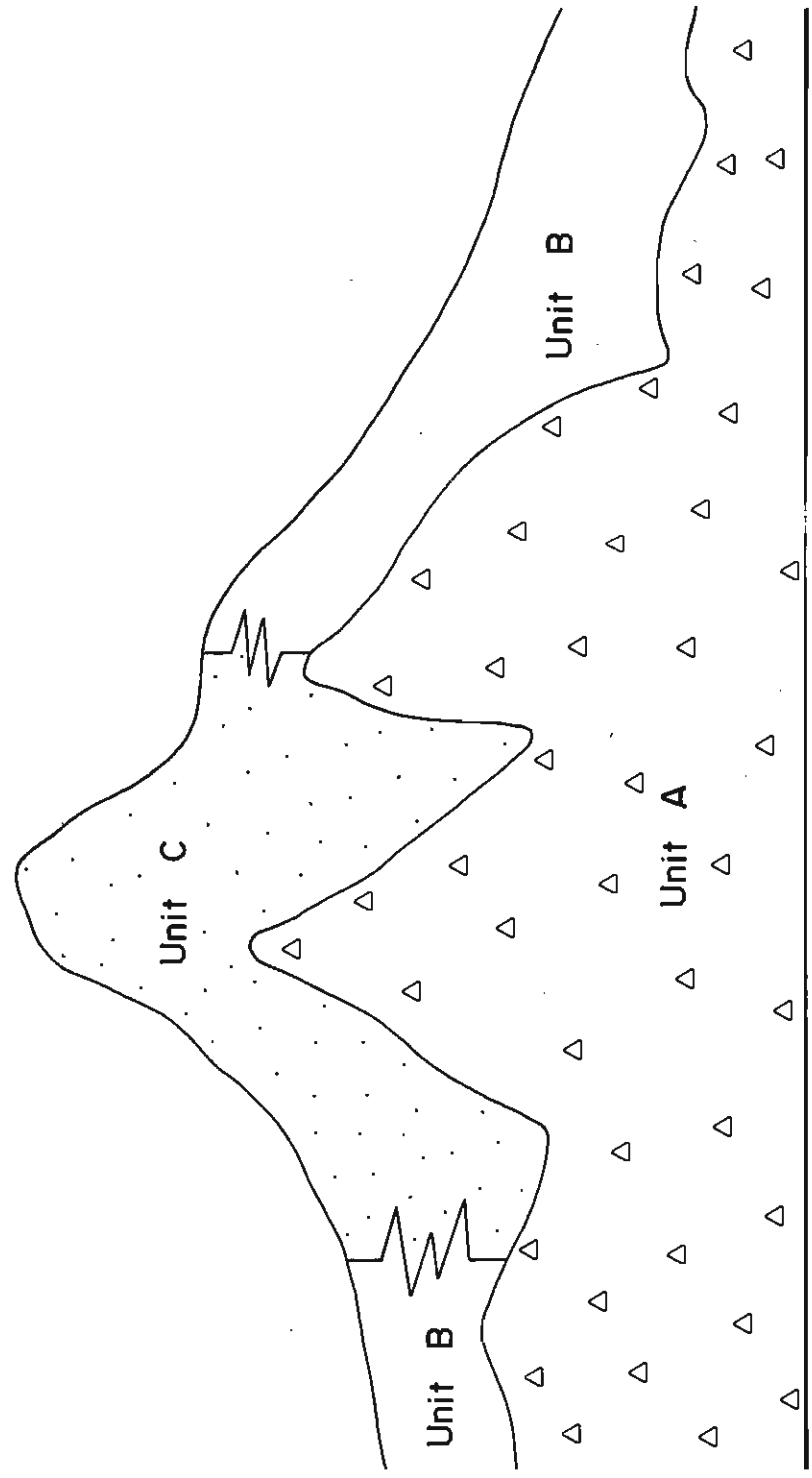
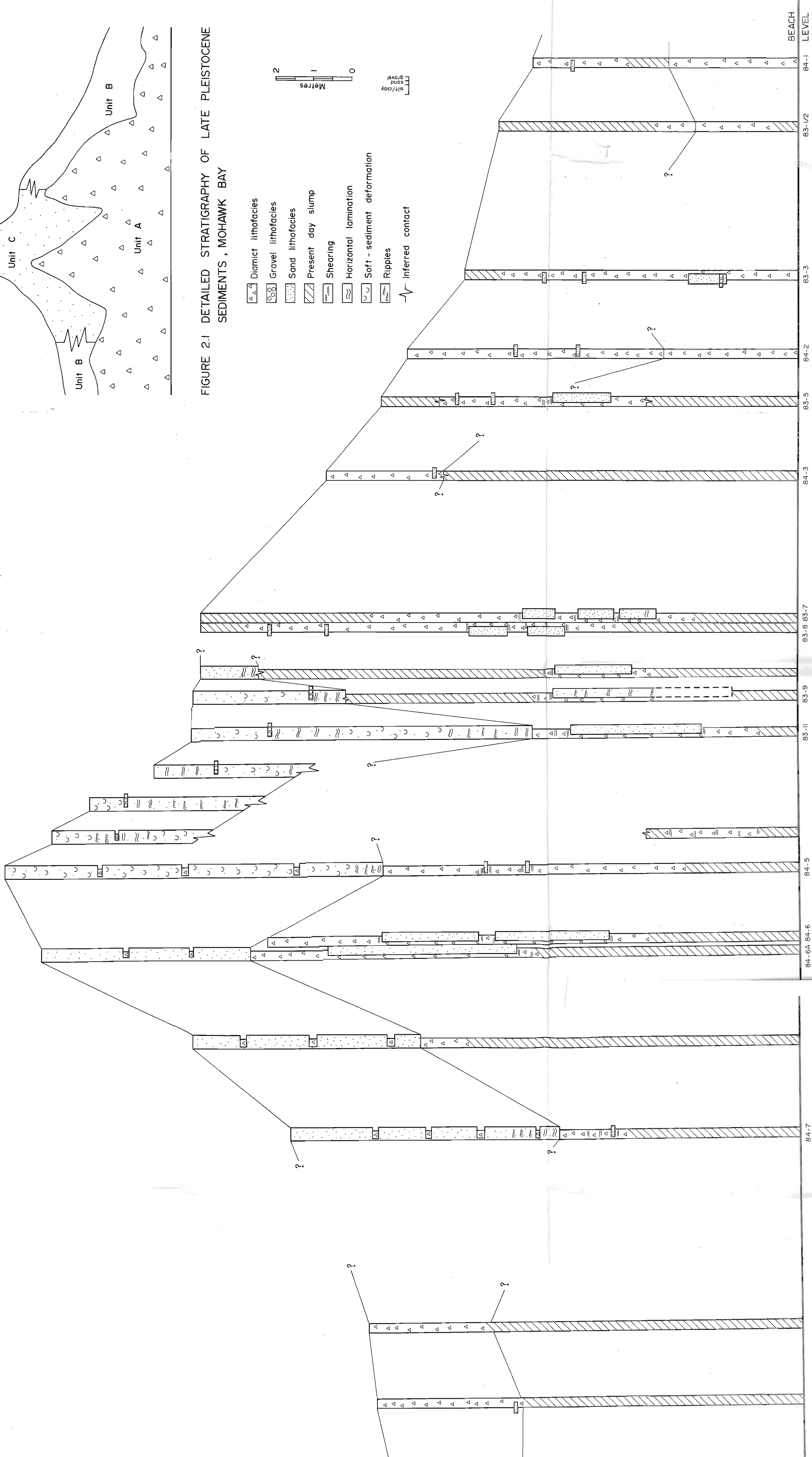
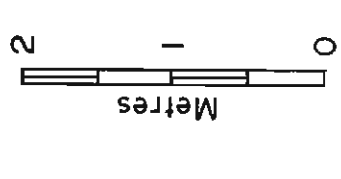


FIGURE 2.1 DETAILED STRATIGRAPHY OF LATE PLEISTOCENE SEDIMENTS, MOHAWK BAY

- Diamict lithofacies
- Gravel lithofacies
- Sand lithofacies
- Present day slump
- Shearing
- Horizontal lamination
- Soft - sediment deformation
- Ripples
- Inferred contact



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